THE UNIVERSITY OF ALBERTA

SOME TERRESTRIAL HEAT FLOW MEASUREMENTS IN ALBERTA

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The undersigned hereby certify that they have read and recommend to the Faculty of Graduate Studies for acceptance, a thesis entitled Some Terrestrial Heat Flow Measurements in Alberta.

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Abstract

Estimates are given of terrestrial heat flow for two Alberta wells, one in the Leduc field and one in the Redwater field, based on temperature measurements in the wells and laboratory determinations of thermal conductivities. Two discrete heat flow regions were observed in the Redwater well, but the deeper was believed to be an anomalous flow region in which the local movement of fluid gave rise to a disturbance of the normal temperatures. For the shallower part of the well, the calculated flow was $1.46 \pm 0.08 \times 10^{-6}$ cal. cm. $^{-2}$ sec. $^{-1}$. This value is accepted as being representative of the undisturbed flow in the well but it is recommended that further measurements be made in wells in the Redwater region to substantiate it.

Calculation of an accurate heat flow for the Leduc well was not possible because of a lack of appropriate conductivity data. Computations based on the limited data available suggest that the flow for this well is as high as, or possibly higher than, that for the Redwater well. The difference may reflect a difference in the radioactive content of the basement rocks underlying the two regions.

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INTRODUCTION

1.1 HISTORICAL. Early observation of such natural phenomena as hot springs and volcanoes led to the belief that the earth's interior is considerably hotter than its surface. The development of deep mines in various localities, in which temperature was found to increase with depth, served to confirm this view. Practical problems, such as the estimation of the depths at which men could work in reasonable comfort, gave rise to the measurement of temperatures in mines and to the calculation of the change in temperature with depth or geothermal gradient. Observations of this type were made as early as 1832 (Brit. Assoc. Advance Sci. 1936) and a considerable body of data is available at the present time (Van Orstrand, 1951). Geothermal gradients range in value from about 5 to about 70 degrees C /km., and are determined by finding the best straight line through a number of temperature-versus-depth points.

In boreholes or mines in which a large number of these points have been determined it often becomes apparent that the temperature-depth relation is not a simple straight line but is best expressed as a series of line segments. When it is possible to correlate the segments with lithology it is found that a change in temperature gradient usually corresponds to a change in lithology. Furthermore, if the thermal conductivities of the geological formations are known, it is observed that an increase in gradient goes hand in hand with a decrease in conductivity. This is precisely what would be expected in the flow of heat through a series of conductors of uniform cross-section but of different conductivities, containing no heat sources or sinks. When the gradient and the conductivity are measured for each formation, the values for terrestrial heat flow so determined are independent of the geological formation.

Although terrestrial heat flow is obviously a much more fundamental quantity than the geothermal gradient, it was not until shortly before the Second World War that any serious efforts were directed towards measuring rock

conductivities in boreholes in which temperatures had been, or were likely to be measured. Interest in conductivity measurement was first generated by the British Association for the Advancement of Science through its Committee on the Conductivities of Rocks (Brit. Assoc. Advance Sci., 1935 et seq.). The work of this committee culminated in the publishing of the first accurate values for terrestrial heat flow by Bullard and Benfield in 1939. Since then a large number of measurements have been made at widely scattered points on the earth's surface. Birch (1954) has summarized the measurements to that date and since then other workers have reported further values. References to terrestrial heat flow determinations since Birch's compilation are listed in the bibliography which supplements the list of references cited. Continental values range from about 0.6 to 3.0 x 10^{-6} cal. cm.$^{-2}$ sec.$^{-1}$. the generally accepted average being about 1.2 x 10^{-6}. There is a much greater variability in the oceanic values (from about 0.1 to 8.0 x 10^{-6}) but the average flow is still believed to be approximately 1.2 x 10^{-6} cal, cm.$^{-2}$ sec.$^{-1}$.

Canadian measurements have been made by Misener and co-workers in the mining areas of Northern Ontario and Quebec (Misener, Thompson and Uffen, 1951) and in a borehole at Resolute Bay, N.W.T. (Misener, 1955). The many oil wells in Western Canada would seem to offer unparalleled opportunities for further Canadian heat flow determinations. It is not uncommon for oil companies to obtain temperature logs (i.e. continuous records of temperature versus depth) when the drilling of a well is completed and core samples are usually obtained from various depths during drilling. Copies of the logs are readily available and it is also possible to get some of the core for conductivity measurements. It is therefore unfortunate that the action of the drill and the circulation of drilling fluids disturb

the normal gradient to the extent that a period of the order of that taken to drill the well is required for thermal equilibrium to be re-established, (Bullard, 1947) rendering oil company logs useless for heat flow determinations.

Temperature Measurements for heat flow purposes must then be made only when sufficient time has elapsed for the normal gradient to be re-established. At the end of this time most wells are either producing and unavailable or have been sealed off in accordance with Government regulations in the case of dry holes. Of the many wells that have been drilled in Alberta, it is not surprising that only a very few are available and suitable for heat flow measurements.

1.2 Geophysical Implications of Terrestrial Heat Flow.

On the basis of earthquake data, the earth is now believed to be made up of a central core, an intermediate layer called the mantle and a relatively thin outer skin or crust. The core, with a radius of about 3400 km., is liquid at least in part, although there is some reason for thinking that the inner core is solid. The mantle and crust are solid. The boundary between the two is called the Mohorovicic discontinuity and is distinguished by an abrupt change in compressional and shear wave velocities. There is some evidence for the existence of another discontinuity of the same kind within the crust in continental areas. This has been called the Conrad discontinuity.

Measurements of the acceleration due to gravity in different parts of the world have given further information about crustal structure. When the necessary corrections to the raw data have been made, a broad pattern emerges which indicates that elevated regions of the earth's surface are associated with reduced gravitational forces, whereas the reverse is true in regions where the earth's surface is depressed. Two hypotheses, due to

Pratt and to Airy, were originally advanced to explain this result. Airy's hypothesis that the crust thickens under mountain ranges and thins in the oceanic areas is now generally accepted. Seismology has confirmed that the crust is thinner under the oceans, the depth to the Mohorovicic discontinuity being about 5 km. as compared to 35 under the continents. The case for a thickened crust under mountain ranges is not as well established.

Petrological analysis of volcanic material in continental and oceanic areas indicates that there is probably a compositional difference between the crusts in the two areas. Continental volcanics are mainly sialic, that is they are predominantly aluminum silicates, whereas oceanic volcanics are predominantly simatic, composed of magnesium silicates. The Conrad discontinuity may correspond to the boundary between sialic and simatic material under the continents.

The existence of a measurable heat flow from the earth's interior raises the problem of its origin. One source which most certainly contributes something to the observed flux is radioactivity. All rock types occurring at the earth's surface in appreciable quantities contain radioactive elements and it is reasonable to assume that this persists to some depth within the earth. The energy released during radioactive decay within the earth is converted into heat energy. Fairly extensive data are available on the radioactive content of typical rocks, so that estimates of the radioactive heat generated in sialic and simatic material may be made. Bullard (1954) estimates the heat production in cal. cm.$^{-3}$ sec.$^{-1}$ to be 53×10^{-14} for sial and 15×10^{-14} for sima. About 20 km. of sial would be sufficient to cause the observed flow of 1.2×10^{-6} cal. cm.$^{-2}$ sec.$^{-1}$. Although some juggling with thicknesses of sialic and simatic material is permissible it appears that if the rocks at a depth are as radioactive as those at the surface, we have no difficulty in explaining continental flow in terms of radioactive

heating alone. Indeed, the problem in this case is explaining why continental flows are not greater. Bullard suggests that there is a continual decrease with depth in the radioactive content of rocks and that surface values represent maxima rather than averages (Bullard, 1954a). He assumes that the average radioactive heating at depth in the crust is about half that calculated for surface rocks. This assumption gives a radioactive heat flow of 0.9×10^{-6} cal. cm.$^{-2}$ sec.$^{-1}$ from the crust, making possible contributions to the observed flow from other sources.

Because the oceanic crust is thinner and is markedly less radioactive than the continental, it was predicted that oceanic flows would be relatively small. Because of the comparitively greater difficulties in measuring oceanic gradients and in obtaining samples of deep-sea sediments it was only in 1952 (Revelle and Maxwell, 1952) that heat flows were published for the Pacific. Bullard published similar values for the Atlantic in 1954 and since then Von Herzen and Maxwell (1959) have published further values for the Pacific. The average flows in both areas are approximately the same as the continental average. Bullard, Maxwell and Revelle (1956) offer two explanations for this somewhat surprising result. According to the first, the continental crust may have been radioactively enriched at the expense of the mantle whereas under the oceans there is the same total amount of radioactivity which is distributed more evenly and to a greater depth.

The second suggestion calls on the hypothesis of subscrustal convection currents, which has been advanced by various authors (Griggs, 1939: Meinesz, 1947) in attempting to explain orogenesis. These authors visualize vast slow-moving convection cells which come into being when unstable temperature conditions are set up in the mantle. If such cells do exist, the normal

heat flow should be enhanced over rising currents and should be less over descending currents. The possibility then arises that, at least under the oceanic areas measured so far, rising currents are bringing up heat. A system of descending currents under the continents may be decreasing the heat flow or the thick continental crust may act as an insulator against such effects. The evidence for or against either of Bullard Maxwell and Revelle's hypotheses is inconclusive at the present time. A definite answer to this and other heat flow problems awaits the accumulation of much more data.

If Bullard's suggestion, that the average radioactivity of crustal rocks at depth is about half that at the surface, is accepted, a certain portion of the observed flow must be ascribed to other sources. Possibilities are energy released in earthquakes or heat brought up by igneous intrusions. Verhoogen (1956) states, however, that the energy involved in these phenomena is a tenth or less of that brought to the surface by terrestrial heat flow.

The last possible heat source depends on the assumption that the earth had a hot-body origin and is now cooling. There is considerable doubt on this score at the present time (Lubimova, 1958) and again an increased mass of heat flow data should help to answer this fundamental question of the earth's origin.

2. DETERMINATION OF TERRESTRIAL HEAT FLOW.

2.1 Steady-state with Horizontal, Homogeneous and Isotropic Layering.

The determination of terrestrial heat flow involves measurements of temperature variation with depth and of the thermal conductivities of the geological formations in which the temperature measurements are made. If the earth is regarded as a homogeneous, isotropic halfspace, containing no near-surface sources or sinks, in which thermal equilibrium has been established, the heat flow near the surface due to deep-seated sources is given by:

$$f = k \frac{dv}{dz} = k\ g \quad \text{.......................................(1)}$$

Where f = heat flow per unit area per unit time.

v = temperature

k = thermal conductivity

z = depth

g = geothermal gradient

Under these conditions the temperature is a linear function of depth and the geothermal gradient is constant.

Usually it is not possible to consider the earth as a homogeneous halfspace. In areas such as Alberta, where some thousands of feet of sediments cover a crystalline basement, the layering due to sedimentation must be considered. If it is assumed that the subsurface consists of a number of horizontal, homogeneous and isotropic layers, the heat flux in the ith layer, perpendicular to the layering, is given by:

$$f = k_i\ g_i \quad \text{..(2)}$$

Where k_i = thermal conductivity of the ith layer

g_i = geothermal gradient in the ith layer

A heat flow value may be computed for each formation encountered in which both temperatures and conductivities have been measured and the mean of these values calculated. Bullard (1939) has, however, indicated a better method of treating the data:

If v_z = temperature at depth z

v_o = temperature at the surface

g_z = geothermal gradient at depth z

k_z = thermal conductivity at depth z

Then
$$v_z = v_o + \int_o^z g_z \, dz \qquad (3)$$
$$= v_o + f \int_o^z \frac{dz}{k_z} \qquad (4)$$

If we now return to the concept of a layered earth
$$v_n = v_o + f \sum_1^n t_i / k_i \qquad (5)$$

Where t_i = thickness of the ith layer

Equation 5 is the analogy for the case of layering of the linear temperature-depth relation which holds when the earth can be considered as a homogeneous halfspace. For the more general case a straight line is obtained if temperature is plotted againt $\sum t_i/k_i$. The slope gives the heat flow and the intercept on the temperature axis gives the extrapolated surface temperature.

The equation of heat conduction in a solid containing no heat sources or sinks is:

$$\nabla^2 v = \frac{1}{\kappa} \cdot \frac{\partial v}{\partial t} \qquad (6)$$

Where $$\nabla^2 \equiv \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2}$$

κ = thermal diffusivity = $k/\rho c$

ρ = density

c = specific heat

t = time

Equation 5 is the steady-state solution of this equation for the case of horizontal, homogeneous and isotropic layering. If the actual geologic conditions cannot be approximated to by this type of layering or if, for some reason, thermal equilibrium has been disturbed and has not been re-established, Equation 5 is no longer the solution of the heat flow equation in which we are interested. We will consider first departures from the steady-state conditions and their effects on the observed heat flow.

2.2 Departures from Steady-state Conditions.

A number of different situations can give rise to temperature disturbances which may be only transient or which may, as far as the geophysicist is concerned, be permanent. If the disturbance is transient it is sufficient to wait for equilibrium to be re-established. If the disturbance is permanent, its effect should be corrected for.

The simplest transient disturbance to deal with is that which is caused by drilling into the subsurface. The sources of temperature disturbance which arise in the drilling of a well are discussed by Guyod (1946). Heat is evolved because of the friction of the drilling bit and the drill pipes and some may also originate because of chemical reactions between the drilling fluid and the formations. The principal cause of temperature

disturbance is, however, the circulation of drilling fluid in the hole, which warms the formations near the surface at the expense of those at the bottom of the hole. Bullard (1947) has calculated that the temperature disturbance in the neighbourhood of a borehole becomes negligible after a period roughly equal to that taken to drill the hole. Temperature measurements in boreholes for the purpose of estimating terrestrial heat flow are generally not made until sufficient time has elapsed for the formations in the immediate vicinity of the hole to regain their original temperatures.

Among the effects which cause departures from steady-state heat flow which are essentially permanent, in the sense that the times required to re-establish thermal equilibrium are short only with respect to the geologic time-scale, are uplift and erosion. Qualitatively, it is easily seen that if the earth's surface is uplifted it is exposed to cooler and cooler regions of the atmosphere. If no temperature discontinuity is to exist at the surface, more heat must flow from the near-surface sections and if the uplift continues over a period of time, this effect penetrates deeper and deeper. Similarly, if erosion is proceeding at some steady rate, the removal of a thin layer from the surface exposes a new surface which is somewhat warmer than the atmosphere with which it is in contact and again a near-surface cooling takes place which penetrates deeper and deeper as erosion continues. If measured temperatures are plotted against $\sum t_i / k_i$, the graph is no longer a straight line but shows an increasing curvature as depth decreases, corresponding to the greater heat flows and larger gradients near the surface. Benfield (1949) has solved the equation of heat conduction in a moving medium and has applied it to examine quantitatively the effects of uplift and erosion (1949a).

Long-period climatic variations also give rise to non-equilibrium conditions. In particular, the advance and retreat of glaciers has caused alternate cooling and warming of the surface. A warming effect will presumably predominate at the present time, if the effect is significant. Since this is the opposite of the effect of uplift and erosion, the action of glaciation should be again revealed in the curve of temperature against $\sum t_i / k_i$ by an increase in curvature with decreasing depth, but in this case there are lesser heat flows and smaller gradients near the surface. Birch (1948) has solved the heat flow equation, assuming sharp, discontinuous temperature changes at the surface as an approximation to the actual changes during past glaciations. His calculations indicate that effects of past glaciations may be felt to depths as great as 10,000 feet. Both Bullard (1939) and Benfield (1939) applied corrections of this type to their results when curvatures were observed in the graphs of temperature plotted against $\sum t_i / k_i$.

2.3 Temperature Disturbance in the Neighbourhood of a Flowing Well.

If a fluid, which may be oil, gas or water, emerges from a formation at some depth below the surface and flows to the surface in a well, the fluid will be cooled and the surrounding strata heated in the process. As the flow continues, the temperature disturbance so generated will penetrate to greater and greater distances from the well. Temperature measurements made in the well fluid will be estimates of the disturbed temperatures in the formations in immediate contact with the well, rather than estimates of the rock temperatures existing before flow commenced.

If the rate of flow is slow enough and if not too much time has elapsed since the flow began, it will be found that above a certain depth

the magnitude of the temperature disturbance remains constant, and the measured geothermal gradient over this range of depths will be identical with the undisturbed gradient. Birch (1947) made temperature measurements in a well in Colorado from which water flowed at the rate of 1.1 gallons per minute. A sudden change in gradient was observed in the immediate vicinity of the supposed inflow depth but the original gradient was restored after a few hundred feet. Birch investigated the region of constant temperature disturbance by assuming that the effect observed was due to a continuous line source of constant strength per unit length (Carslaw and Jaeger, 1959), and he was able to deduce a value for the conductivity of the formations through which the well passed. He combined this with the value of the geothermal gradient to get an estimate of the heat flow.

If the rate of fluid flow is sufficiently fast, or if a long time has elapsed since flow began, the magnitude of the temperature disturbance will be found to vary continuously with depth. Boldizsar (1958) developed an equation expressing the disturbance as a function of the vertical distance from the point of inflow, which is useful in this more general case:

$$v_f - v_r = \frac{V\rho_f c_f}{kF(\psi)} \cdot g \left[1 - \exp.\left\{ \frac{-kF(\psi)}{V\rho_f c_f}(z_f - z) \right\} \right] \qquad (7)$$

Where v_f = Temperature of the fluid i.e. temperature of the rock in contact with the well.

v_r = Rock temperature before flow started.

V = Volume flow rate of the fluid.

ρf = Density of the fluid.

c_f = Specific heat of the fluid.

z_f = Depth at which fluid flows into the well

$\psi = \kappa t / a^2$

κ = Thermal diffusivity of the rock.

a = Radius of the well.

The exact nature of the function $F(\psi)$ is dependent on the solution of the heat flow equation which is assumed to hold for the flow of heat from the fluid in the well into the surrounding formations. Boldizsar has made use of Jaeger's solution for the flow of heat in the region bounded internally by a circular cylinder (1942). If the surface of the cylinder is kept at constant temperature $(v_f - v_r)$ and the region outside the cylinder is initially everywhere at zero,

$$F(\psi) = \frac{8}{\pi}\int_0^\infty \frac{e^{-\psi u^2}}{J_0^2(u) + Y_0^2(u)} \cdot \frac{du}{u} \qquad \ldots\ldots\ldots\ldots\ldots(8)$$

Where u = Variable of integration.

J_0 = Zero order Bessel function of the first kind.

Y_0 = Zero order Bessel function of the second kind.

Values of $F(\psi)$ as given by Equation 8 are tabulated by Ingersoll, Zobel and Ingersoll (1954).

Although it is possible to use Boldizsar's $F(\psi)$ in investigating the effects of pumping on subsurface temperatures, we will want also to determine the magnitude of the temperature disturbance in wells which have been shut down, and have, therefore, had an opportunity to regain some measure of thermal equilibrium. For this purpose it is more convenient to determine the form of $F(\psi)$ for a continuous line source.

The temperature v at a radial distance r from a continuous line source of strength q per unit length is (Carslaw and Jaeger, 1959):

$$v = \frac{q}{4\pi K}\int_{r^2/4\kappa t}^{\infty} \frac{e^{-u}\,du}{u} = \frac{-q}{4\pi K}\, Ei\left(\frac{-r^2}{4\kappa t}\right) \qquad \text{..........(9)}$$

Where t = time since the initiation of the source.

$Ei(-u)$ = exponential integral

$$\frac{\partial v}{\partial r} = \frac{-q}{2\pi K}\cdot\frac{e^{-r^2/4\kappa t}}{r} = \frac{2}{r}\left[\frac{e^{-r^2/4\kappa t}}{Ei(-r^2/4\kappa t)}\right]v$$

$$\left(\frac{\partial v}{\partial r}\right)_{r=a} = \frac{2}{a}\left[\frac{e^{-a^2/4\kappa t}}{Ei(-a^2/4\kappa t)}\right]v(a,t) \qquad \text{.........(10)}$$

Let radial heat flow per unit length be f

$$f = -2\pi k a\left(\frac{\partial v}{\partial r}\right)_{r=a}$$

$$= 4\pi k\left[\frac{e^{-a^2/4\kappa t}}{-Ei(-a^2/4\kappa t)}\right]v(a,t)$$

$$= 4\pi k\left[\frac{e^{-1/4\psi}}{-Ei(-1/4\psi)}\right]v(a,t) \qquad \text{...........(11)}$$

Following Boldizsar's development, the heat flowing out through a length dz of the well in unit time is related to the temperature drop in the fluid over the same length by the following equation:

$$dv_f = \frac{+f\,dz}{V\rho_f c_f} \qquad \text{............(12)}$$

Substituting for f from Equation 11:

$$dv_f = \frac{+4\pi k}{V\rho_f C_f}\left[\frac{e^{-1/4\psi}}{-Ei(-1/4\psi)}\right](v_f - v_r)\,dz$$

$$= \frac{+kF(\psi)}{V\rho_f C_f}\left\{v_f - v_0 - gz\right\}dz \quad \text{..............(13)}$$

The solution of Equation 13 with the boundary condition $v_f = v_r$ at $z = z_f$ is Equation 7. $F(\psi)$ in the case of the line source is:

$$F(\psi) = 4\pi\left[\frac{e^{-1/4\psi}}{-Ei(-1/4\psi)}\right] \quad \text{..............(14)}$$

When ψ is very large (i.e. t large and/or a small):

$$F(\psi) \simeq \frac{1 - a^2/4\kappa t}{-\gamma - \ln(a^2/4\kappa t)}$$

$$= \frac{1}{-\gamma - \ln(a^2/4\kappa t)}$$

Where γ = Euler's constant ....(15)

This approximation is valid in the problem being considered except when the time since pumping began is very short.

Substitution of the appropriate values for the constants in Equation 7, using $F(\psi)$ as defined by Equation 15, allows the effect of fluid flow on formation temperatures in the vicinity of the well to be calculated for any time during the period when the well is in production. If the well is a gas well, or is producing oil containing appreciable quantities of gas, a further source of temperature disturbance must be considered. The gas in the producing formation is at a pressure considerably

higher than atmospheric and as it enters the well and rises to the surface it expands and cools the surrounding formations. If the pressure in the formation is appreciably greater than that in the well at the same depth due to hydrostatic pressure, most of the expansion will take place as the gas enters the well. Oosterkamp (1948) developed a solution of the heat conduction equation which may be applied to the calculation of the temperature disturbance due to the expansion of gases entering the well. If heat is supplied at the rate q per unit time per unit area for $t > 0$ over the circle $r < a$, $z = 0$, the temperature at the point (0, 0, z) is:

$$v = \frac{2q(\kappa t)^{\frac{1}{2}}}{k}\left\{ \mathrm{ierfc}\,\frac{z}{2(\kappa t)^{\frac{1}{2}}} - \mathrm{ierfc}\,\frac{(z^2+a^2)^{\frac{1}{2}}}{2(\kappa t)^{\frac{1}{2}}} \right\} \qquad \ldots\ldots\ldots\ldots(16)$$

Where $\mathrm{ierfc}\,\beta = \int_\beta^\infty \mathrm{erfc}\, u\, du$

$$\mathrm{erfc}\,\beta = \frac{2}{\sqrt{\pi}} \int_\beta^\infty e^{-u^2}\, du$$

Values of ierfc β are tabulated by Kaye (1955).

The effect of expansion as the gas rises to the surface may be treated by assuming a continuous line source.

It is of interest also to determine the magnitude of the residual temperature disturbance after a producing well has been shut down. It may be assumed that at the time of shutdown the line source continues to function, but that its effect is reduced by the action of a continuous line sink of equal strength. This approach has been used with success in an analogous hydrologic problem (Theis, 1935), the return to equilibrium of a well in a confined aquifer after a period of pumping. Using this

concept of a simultaneously acting source and sink, the temperature of the rock in contact with the well at time t is given by:

$$v = \frac{-q}{4\pi K}\left[Ei\left(\frac{-a^2}{4Kt}\right) - Ei\left(\frac{-a^2}{4Kt'}\right)\right] \qquad \text{.............(17)}$$

Where t' = time since the well was shut down.

t = T + t'

T = total time well was pumped.

When t = T

$$v_T = \frac{-q}{4\pi K} \; Ei\left(\frac{-a^2}{4KT}\right)$$

$$-\frac{q}{4\pi K} = \frac{v_T}{Ei(-a^2/4KT)} \qquad \text{.............(18)}$$

Substituting into Equation 17:

$$\frac{v}{v_T} = \frac{Ei(-a^2/4Kt) - Ei(-a^2/4Kt')}{Ei(-a^2/4KT)}$$

$$\simeq \frac{\gamma + \ln(a^2/4Kt) - \gamma - \ln(a^2/4Kt')}{\gamma + \ln(a^2/4KT)}$$

$$= \frac{\ln(t'/t)}{\gamma + \ln(a^2/4KT)}$$

$$= \frac{\ln(t/t')}{\ln(4KT/a^2) - \gamma} \qquad \text{.............(19)}$$

Equation 19 may be used to determine the residual temperature disturbance in the rocks in contact with the well at any time t' after the well has been shut down.

2.4 Effects of Topography on Heat Flow.

If steady-state flow has been established but geologic conditions are such that the layering may not be considered to be homogeneous, horizontal and isotropic, the isothermal surfaces will be warped and the heat flux will be a function of position. Information about the sub-surface is usually inadequate to allow any corrections to be made for non-ideal layering but the special case of irregular surface topography is easily corrected for. Jeffreys (1938) has developed a theory for the correction of observed gradients when measurements are made in regions of irregular topography, and Bullard (1938) has applied the theory to heat flow measurements both in the Alps and in more level terrain. He states that except in mountainous country the correction is less than 3%. The calculated correction for some temperature measurements in the Simplon Tunnel was 14%.

Birch (1950) has considered the effect of irregular topography in conjunction with those of uplift and erosion in connection with heat flow determinations made in the Adams Tunnel in the Front Range in Colorado. In this case terrain effects alone increased the heat flow value by about 20% but the effects of uplift and erosion caused reductions which varied from about 7 to 15%, depending on the assumptions made about the geologic history of the region.

3. GEOLOGY OF THE REDWATER AND LEDUC AREAS.

The Alberta plains in the Redwater and Leduc areas are underlain by some six to nine thousand feet of sedimentary rock, beneath which lies the Precambrian crystalline basement. Regional dip in the sedimentary strata is to the southwest. Because of physical limitations in instrumentation, the temperature measurements described in this study were confined to depths of 3000 feet and less. The formations at these depths are mainly Cretaceous in age but some Devonian formations are encountered in the Redwater area, where the effect of regional dip brings the older strata closer to the surface. In Table 1 are listed the Cretaceous and Devonian formations of interest. The data in the table are taken from the Schedule of Wells Drilled for Oil and Gas in 1957, published by the Alberta Oil and Gas Conservation Board (This annual publication is referred to hereafter as the Schedule of Wells).

Shaw and Harding (1954) provide additional data on the Lea Park and Belly River formations. Their analysis is based on the correlation of electric logs from selected wells across the province. In eastern Alberta the Belly River formation is found to consist of an interfingering succession of marine shales and deltaic sands but the formation appears to be undivided in the Leduc area and for some fifty miles or more to the east. The character of the formation in the Redwater area is open to conjecture, but it occurs so close to the surface in this case that the question is, for our purposes, academic. Shaw and Harding describe the undivided formation as a "series of gray to brownish-gray to greenish-gray to gray, carbonaceous shales and silts. Thin carbonaceous layers are characteristic of the normal facies".

Table 1

Near-Surface Geologic Formation in the Leduc and Redwater Fields.

Name	Geologic Age	Description
Edmonton	Cretaceous	Medium grey to brownish and greenish grey shale, sandy shale, sandstone (calcareous to bentonitic), thin beds of bentonite, nodules and thin beds of ironstone, coal.
Belly River	Cretaceous	Grey sandstone alternating with grey and greenish-grey shale. Includes the Oldman, Birch Lake, Grizzly Bear and Ribstone members.
Lea Park	Cretaceous	Dark brownish grey to blue-grey shale and sandy shale, ironstone nodules.
Colorado	Cretaceous	Dark grey bentonitic shale with some thin silt or sand lenses and ironstone nodules.
Second white speckled shale	Cretaceous	Marker bed in the Colorado group. Dark grey shale with minute specks of cream and buff colored calcareous material.
Viking	Cretaceous	Marker bed in the Colorado group. Series of dark grey shales, interbedded with fine to medium grained cherty and glauconitic sandstones.
Blairmore	Cretaceous	Subdivided into the coally series, containing interbedded sands, silts, shales, siderite, coal and carbonaceous beds; the glauconitic-ostracod series, containing glauconitic sandstone and shale, fossiliferous shale and earthy to calcareous quartz sandstone; and the quartz sand series locally containing carbonaceous to coally lenses near the base. The last series is also known as the Ellerslie member.
Wabamun	Devonian	Buff and grey limestones and dolomites, sometimes porous; anhydrite.
Calmar	Devonian	Greenish grey quartz silt. Member of Winterburn group.
Nisku	Devonian	Dolomite, silty dolomite, anhydrite, often porous and locally producing oil.
Ireton	Devonian	Green argillaceous dolomite and green calcareous shale.

4. TEMPERATURE MEASUREMENT TECHNIQUES

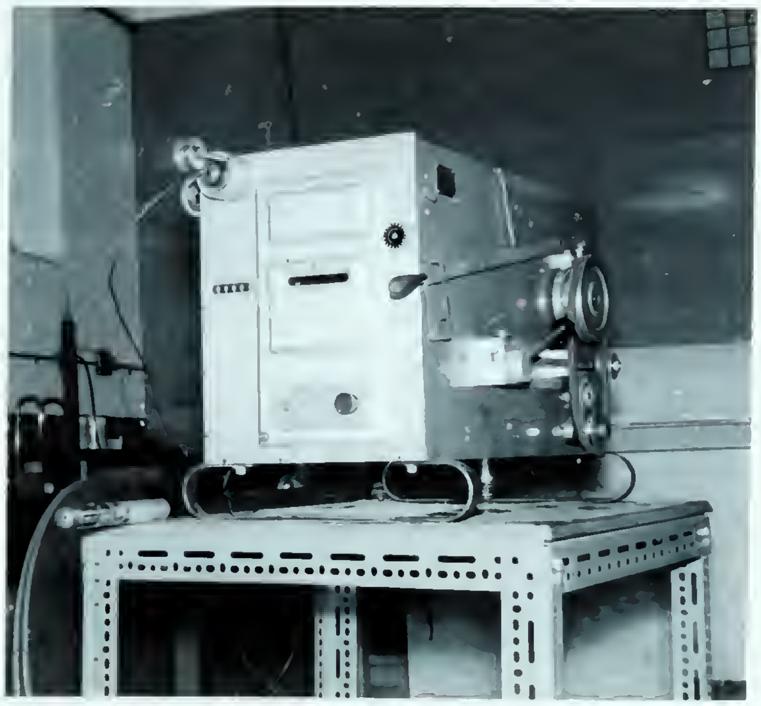
4.1 General

In order that continuous records of temperature against depth might be obtained a temperature logger (Figs. 1 and 2) was acquired. This unit, manufactured by Well Reconnaissance Inc. of Dallas, Texas is known as a "Geologger".

A reel mounted inside the unit carries 3000 feet of two-conductor cable, to the end of which a plummet, or probe, is attached. The probe contains four thermistors connected in series-parallel. The thermistors, which form the temperature-sensing element, are electrically connected through the cable to the measuring circuit on the surface. In the measuring circuit thermistor resistance is converted to voltage and the voltage causes a deflection of a meter needle. An optical system casts a shadow of the needle through a slit upon a light-sensitive ozalid paper. As the probe is raised in the well, the ozalid paper, which is preprinted with footage lines, is fed past the slit into a chamber containing a pair of ammonia-soaked felt wicks. A few minutes is allowed to elapse after the probe has reached the surface and then a complete temperature-depth record, or log, may be removed from the developing chamber. Power for the measuring circuit and to pull the cable and probe up to the surface is supplied by a Homelite 60-cycle, 115-volt gasoline generator.

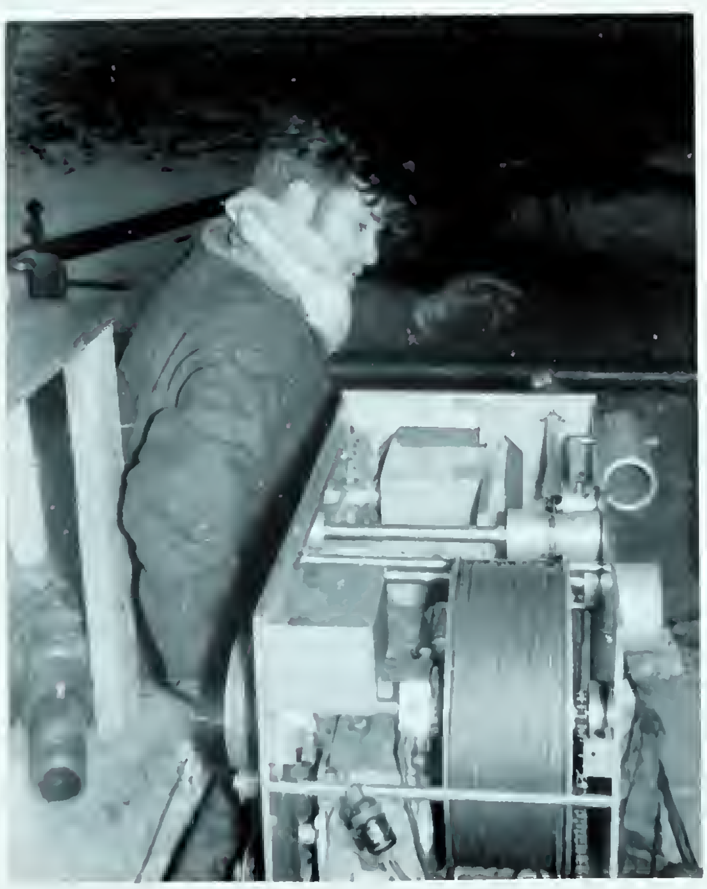
The thermistors are protected from damage by a metal cage which allows the fluid in the well to circulate past them, and which screws on the bottom of the probe. Since their accuracy was unknown, it was decided that it would be advisable to be able to spot check the continuous temperature logs and for this purpose an additional cage was constructed to fit between the probe and the thermistor cage. The cage was designed to hold

FIGURE 1



FRONT VIEW OF TEMPERATURE LOGGER

FIGURE 2



REAR VIEW OF TEMPERATURE LOGGER

either of two types of maximum thermometer, both supplied by G.H. Zeal of London, England. Type AO has a temperature range of -38 to 120°F but its readings are subject to considerable error at pressures appreciably above atmospheric, such as the hydrostatic pressures existing at depth in fluid-filled wells. Type AP is especially constructed to give accurate readings under high-pressure conditions but its temperature range is only from 80 to 120°F. Because of their special construction, type AP thermometers have relatively high heat capacities and require about 10 minutes to reach thermal equilibrium compared to about 3 minutes for AO thermometers.

Early in the investigation serious discrepancies were observed between the temperature logs and spot temperatures taken at various depths with the maximum thermometers. Log temperatures were consistently 20 to 30°F higher than thermometer temperatures. Although some of this effect may have originated in leakage of well waters into the probe, causing partial shorts across the thermistors, the major part was brought about by heating of the thermistors due to passage of the measuring current. This defect has not been remedied and the temperatures tabulated in this report were all taken by maximum thermometer.

4.2 Pressure Correction to Type AO Thermometers, Well Constants Known and no Fluid Flow in Well.

Because of the limited range of the type AP thermometers, they can only be used for temperature measurements in Alberta wells at depths where the formation temperatures are 80°F and above. These depths are some 1200 feet below the fluid level and if temperature observations are to be obtained over this interval, they must be made with the AO thermometers, which will give results which will be more or less in error, depending on the hydrostatic pressure. It can be shown that the hydrostatic

pressure p_z at a depth z below the fluid level in a well with a uniform geothermal gradient g is given by:

$$p_z = \rho_R \left[1 - \alpha(v_0 - v_R)\right] z - \frac{\rho_R \alpha g}{2} z^2$$

.............(20)

Where ρ_R = density of well fluid at some reference temperature v_R.

α = coefficient of volume expansion of well fluid.

v_o = temperature of fluid at fluid surface.

This holds for an incompressible fluid. If the fluid is compressible another expression showing the dependence of p_z on the compressibility may be developed but the effect is smaller than that described by equation 20 and we may neglect it.

For the depths and wells considered in this report, the quadratic term in the equation may also be neglected and the variation of pressure with depth is very nearly linear. Under these conditions the pressure at any depth is primarily a function of the location of the fluid level and the fluid density, and it may be calculated if these two quantities are known. If, in addition, the AO thermometers used have been calibrated so that the pressure effect is known, corrected temperatures may be obtained for any depth. Calibration of a pair of these thermometers is described in Appendix I.

In the discussion above it has been tacitly assumed that the apparent temperature increase due to pressure is independent of temperature. This temperature increase is actually a consequence of a decrease in bulb volume and Busse (1941) states that the decrease in volume of a cylindrical thermometer bulb due to external hydrostatic pressure is directly proportional

to the bulb diameter and inversely proportional to the wall thickness. Both of these should be nearly independent of temperature and we would expect, therefore, that the pressure-induced temperature increase would be also. Appendix I presents data to support this view.

4.3 Pressure Correction to Type AO Thermometers, Well Constants Unknown or Fluid Flowing in the Well.

Direct computation of the pressure correction to the type AO thermometers depends on a knowledge of fluid level and density. If the well has been a producer the oil company concerned can usually supply a figure for the density. Location of the fluid level presents more of a problem. Its value can fluctuate over a period of time, and if direct computation is to be relied on, the fluid level must be checked frequently, at least until the magnitude of the variation has been ascertained. Location of the fluid level is a tedious procedure and can involve the sacrifice of time better spent in obtaining temperature data.

It is sometimes desirable, therefore, that an empirical method of correcting for the pressure effect be found. Occasionally the use of an empirical method may be mandatory. Equation 20 is valid if the fluid in the well is everywhere static. It is possible, however, for fluid to move into the well from an adjacent formation through a corroded section of casing. The fluid may then flow through the well and eventually escape into another exposed formation. If this is the case, the pressures existing in the regions of flow will not be those given by equation 20, but will be less by amounts depending on the volume rate of flow. Generally, empirical corrections in the case of fluid flow within a well will not interest us, since the regions of flow will also be regions of temperature disturbance.

We will be unable to correct our temperatures to the undisturbed values because we will usually have no knowledge of the volume of flow or the time since it began.

For depths for which formation temperatures are 80°F and higher, temperature readings may be obtained at every depth with both types of thermometer. Since the pressure effect is temperature-independent and pressure is a linear function of depth, the temperature difference between the two thermometers should be a linear function of depth also and the intercept on the depth axis should give the fluid level. If a straight line is obtained its slope should give a reasonable estimate of the density for the well fluid and its intercept a reasonable estimate of the fluid level. If these values are not reasonable, or if the plot is not linear, some sort of pressure disturbance should be suspected. If the values are reasonable, the straight line will furnish a basis for correcting the type AO thermometer readings for pressure.

4.4 Adjustment of Type AO Thermometer Readings for Individual Thermometer Differences.

During the period when well temperatures were being measured, two of the type AO thermometers were used. Comparisons of the readings given by the two when simultaneously immersed in a cooling bath had been previously made in the laboratory. On the basis of the results obtained the temperatures given by thermometer A02899 have been adjusted so they would be directly comparable with those obtained with A02906 with which most of the temperature readings were taken.

5. Thermal Conductivity Measurements.

5.1 Divided Bar Apparatus.

For the determination of thermal conductivities a divided bar method, similar to that described by Benfield (1939), is used. The apparatus consists of a pair of cylindrical brass bars, 1 inch in diameter and 4 inches in length, between which discs of rock, also 1 inch in diameter and of varying thicknesses, may be placed. The bars are mounted vertically with the rock sample in place; the upper bar surmounted by another brass cylinder, of the same diameter and 1 inch in length, containing a heater; the lower with the bottom end surrounded by a coil through which cooling water may be circulated. Wood insulating blocks are placed above the heating unit and below the lower brass bar and the whole assembly is held rigid in an hydraulic press.

When the heater is switched on and the cooling water begins to circulate, heat flows through the apparatus and temperature gradients are set up in the brass bars. Determination of the thermal conductivity of the sample involves the measurement of the steady-state temperature distribution in the bars. For this purpose 4 copper-constantan thermo-couples are permanently cemented into holes in each bar. The holes are equally spaced at 1-inch intervals along the length of the bars and the two next to the rock sample are just 3 mm. from the ends of the bars. Water and heater temperatures are measured with mercury thermometers and the flow of heat is adjusted to bring the average of the two (i.e. the sample temperature) as close to room temperature as possible. If this is done, radiation losses are reduced to a minimum. Temperatures within the brass bar are estimated from the measurement of thermo-electric voltages between pairs of thermo-couples.

If there are no radiation losses, the heat flow through all cross-sections of the apparatus will be the same. If the subscripts r and b refer to rock and brass respectively, and f, k and g have their usual significance, we may write:

$$f = k_r g_r = k_b g_b \qquad \text{.......................................(21)}$$

Let v = temperature drop between the thermocouples nearest to the rock sample.

a = distance of thermocouple from end of brass bar.

s = sample thickness.

L = temperature drop across the two rock-brass interfaces due to thermal contact resistance.

Then $v = 2a\, g_b + L + g_r s$

$$= M\, g_b + g_r s \qquad \text{.......................................(22)}$$

The introduction of the expression $M\, g_b$ is based on the interfacial temperature drop being proportional to the heat flux. M may be regarded as the thickness of brass over which there is a temperature drop of $2a\, g_b + L$.

Dividing through by g_b and substituting from equation 21:

$$v/g_b = M + (k_b/k_r)\, s \qquad \text{.......................................(23)}$$

If the value of the interfacial temperature drop were known the conductivity of a single rock sample could be derived from equation 23. Unfortunately, the magnitude of the interfacial drop depends on the sample itself. For this reason it is necessary for measurements to be made on a number of samples of the same rock with differing thicknesses. If the various discs have been prepared in the same way, the interfacial temperature drops will be the same for all and if v/g_b is plotted against the sample thicknesses, a straight line will be obtained with intercept M and with slope the ratio of the conductivities in the rock and brass.

5.2 Calibration of the Divided Bar Apparatus

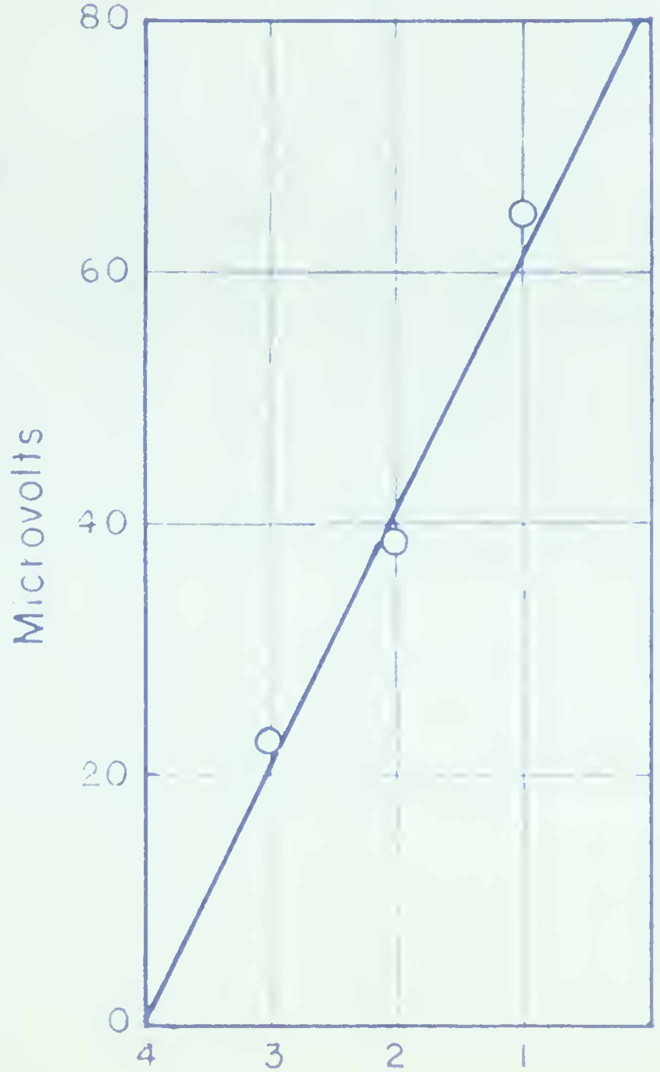
The determination of accurate rock conductivities depends on an accurate knowledge of the conductivity of the brass bars. Because the conductivity of brass varies considerably with the proportions of copper and zinc, the conductivity of the bars must be determined experimentally. To do this, a material of accurately known conductivity is substituted for the rock samples. The material chosen in this study was crystalline quartz. Four discs were made up with thicknesses of 3.02, 5.98, 9.03, and 12.03 mm. The discs were cut so that the plane faces were perpendicular to the optic axis. The most recent measurements of the thermal conductivity of crystalline quartz parallel to the optic axis were made by Birch and Clark (1940). During the calibration of the brass bars the estimated temperatures of the quartz discs varied between 21.5 and 27.4°C. According to Birch's and Clark's results this corresponds to a variation in the thermal conductivity between .0250 and .0244 cal. $sec.^{-1}$ $deg.^{-1}$. $cm.^{-1}$ The value .0247 was adopted for the calculation of the conductivity of the brass bars.

Two measurements were made on each quartz disc, with the exception of the 3.02 mm. disc, on which three measurements were made. The determination of g_b for one of the discs is shown on figure 3 and the graph of v/g_b against disc thickness is plotted on figure 4. A least-squares straight line was calculated for the points. The intercept was 62.0 mm. and the slope 10.45. The conductivity of the brass is, therefore:

$$k = 10.45 \times .0247 = .258 \text{ cal. sec.}^{-1} \text{ cm.}^{-1} \text{ deg.}^{-1}$$

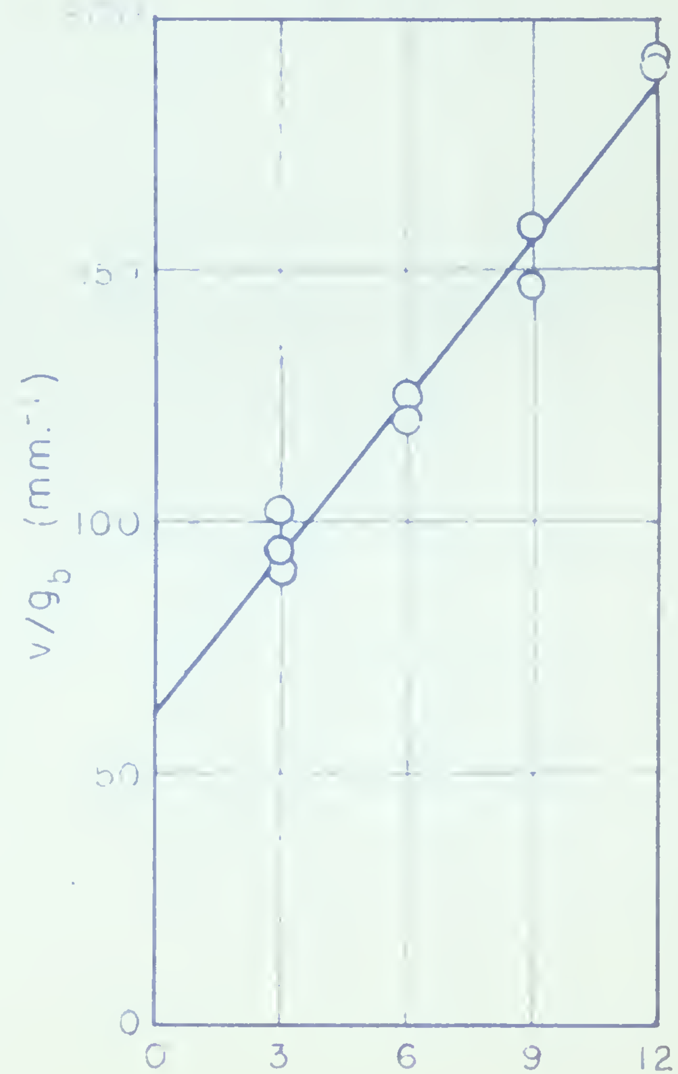
The standard deviation in the slope was 0.54 and in k was .013.

FIGURE 3



DETERMINATION OF g_b

FIGURE 4



DETERMINATION OF BRASS CONDUCTIVITY

5.3 Factors Affecting Validity of Conductivity Results.

The physical conditions to which a sample is subjected when its conductivity is measured in the laboratory are usually quite different from those existing in the formation from which the sample was taken. When a core sample is brought to the surface the principal changes it experiences are a release of pressure, a decrease of temperature and, unless special precautions are taken, a loss of its contained moisture. Each of these factors causes a corresponding change in the thermal conductivity.

Bridgman (1931) has investigated the effect of high pressures on the conductivities of a number of minerals. Assuming that the pressures existing in subsurface formations are roughly the same as the hydrostatic pressures in a column of water extending downwards from the surface, Bridgman's results indicate that the maximum percentage decrease in conductivity due to removal of the sample to the surface would be about one half of one percent. Birch (1942) also presents some data on the pressure effect. Birch's figures suggest the decrease may be as large as 3 percent.

The effects of temperature changes on measured conductivities have been determined by Birch and Clark (1940) for a variety of typical subsurface materials. For sedimentary rocks, the temperature change due to removal of core sample from a depth of 3000 feet can result in an increase in thermal conductivity of about 6 percent.

Birch (1942) also presents results on the effects of wetting on thermal conductivity. The drying of a sample on removal from the formation can cause a decrease in conductivity of up to 30 percent. Because this is such a large source of error, attempts were made to saturate the samples before measurement. The degree of saturation was uncertain, however, and

it was observed that the samples dried out during the conductivity measurement. No effort was made to reproduce formation temperatures and pressures.

5.4 Conductivity Results.

A number of core samples were obtained from co-operating oil companies and from the Oil and Gas Conservation Board in Calgary. It was impossible to get core from the wells in which temperature measurements were made, and difficult to get it, for all formations encountered, from wells nearby. Consequently, the wells from which core was obtained, are, in some cases, considerably removed from the general area in which temperatures were measured (Fig.5).

In preparing the samples for conductivity measurements, a central cylindrical portion, 1 inch in diameter, was cut from each core and discs of various thicknesses were cut from each cylinder. The plane faces of the discs were then smoothed and polished and the thickness of the discs controlled by insisting that the average thickness for four measurements around the edge of the disc not differ by more than .0002 inches from the thickness measured at the center.

The conductivity results are listed in table 2. Also listed are the well numbers corresponding to the well locations shown on figure 5 (well names in full are given in Appendix II), sample descriptions and the formations from which they were taken, and the calculated depths at which the conductivities apply in the two wells in which temperature measurements were made. The calculated depths are based on a comparison of the lithologic logs for the cored wells with those for Leduc 502 and Egremont 42.

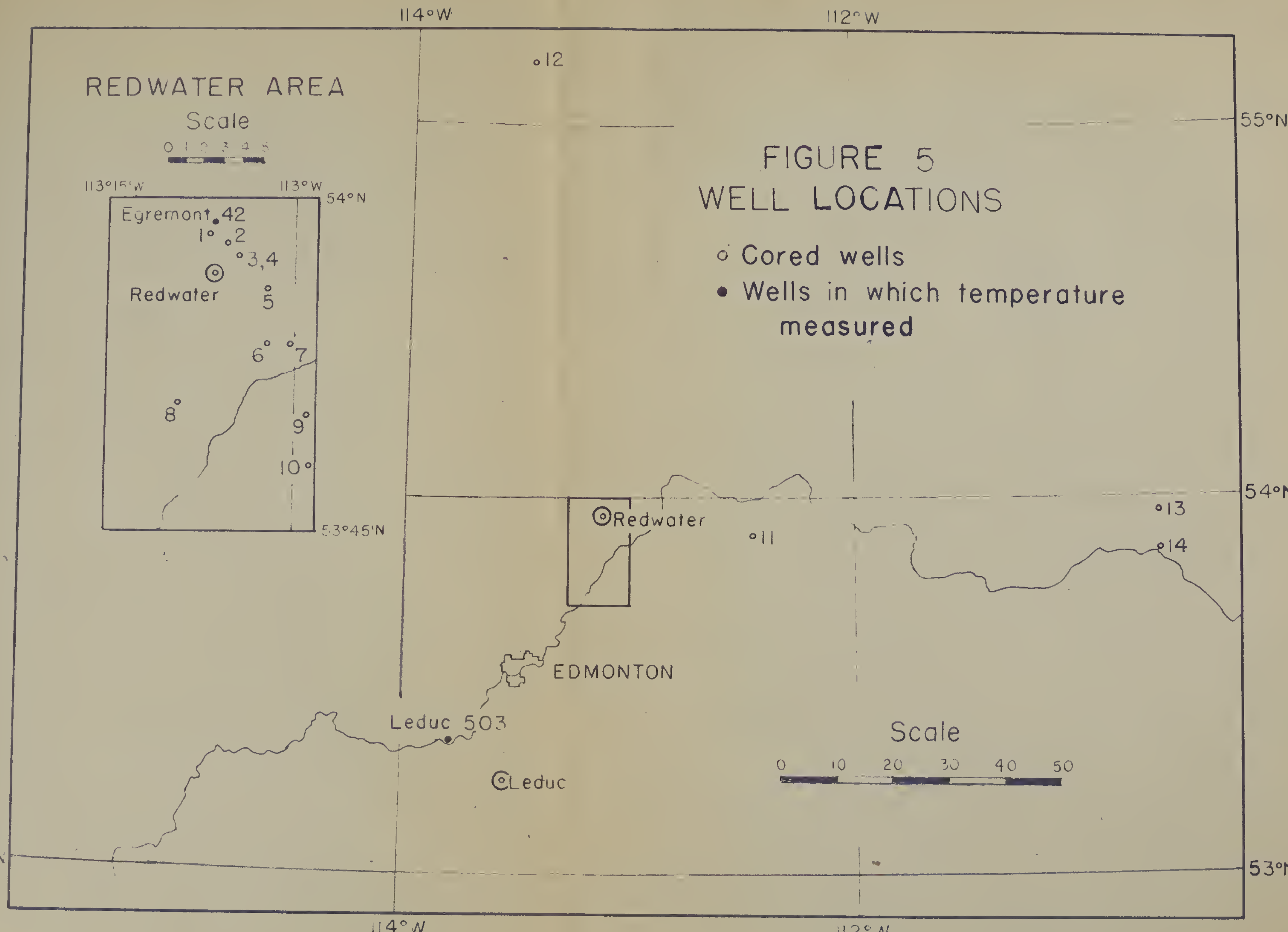


FIGURE 5
WELL LOCATIONS

Table 2

Thermal Conductivities

Sample No.	Well No.	Formation	Sample Description	Equivalent Depth (feet) in Imperial Leduc No. 503	Equivalent Depth (feet) in Imperial Egremont No. 42	Conductivity (cal. sec.$^{-1}$ cm.$^{-1}$ deg.$^{-1}$)
1	14	Colorado	Dense, hard shale	2264	1134	.0037
2	4	2nd white	Dense shale	3119	1835	.0024
3	7	specks	Dense shale	3217	1955	.0047
4	12		Dense hard shale	etc.	1955	.0026
5	1		Dense, hard shale		1962	.0049
6	6		Dense shale		1963	.0044
7	10		Dense, hard shale		1965	.0058
8	1	Viking	Slightly silty shale		1992	.0096
9	11		Soft, brittle shale		1999	.0030
10	1		Silty shale		2010	.0121
11	1		Silty shale		2041	.0056
12	1		Soft, carbonaceous shale		2050	.0060
13	1		Somewhat silty shale		2056	.0024
14	6		Dense shale		2058	.0052
15	4		Soft, carbonaceous shale		2070	.0022
16	11	Blairmore	Medium-to-fine-grained Sandstone		2112	.0077
17	13		Very silty shale		2188	.0031
18	11		- - - - - - - - - - - - - -		2261	.0040
19	11		Soft, brittle shale		2304	.0038
20	10		Shale-sandstone mixture		2314	.0044
21	9		Slightly silty shale		2482	.0035
22	5		Soft, brittle, medium-grained sandstone		2548	.0031

Table 2 (cont.)

Thermal Conductivities

Sample No.	Well No.	Formation	Sample Description	Equivalent Depth (feet) in Imperial Leduc No. 503	Imperial Egremont No. 42	Conductivity (cal. $sec.^{-1}$ $cm.^{-1}$ $deg.^{-1}$)
23	1	Glauconitic	Dense shale		2559	.0089
24	1	Sandstone	Dense shale		2569	.0053
25	1		Very silty carbonaceous SHALE		2578	.0153
26	8		Hard carbonaceous shale		2583	.0060
27	1		Medium-grained sandstone		2586	.0082
28	8		Dense, hard shale		2592	.0117
29	8		Dense shale		2600	.0100
30	8		Fine-grained sandstone		2610	.0118
31	8		Dense shale		2618	.0151
32	10		Dense shale		2634	.0055
33	1		Hard shale		2653	.0066
34	1		Hard, slightly silty shale		2659	.0086
35	11		Fine-grained sandstone		2673	.0040
36	11		Coal		2729	.0022
37	11		Medium-grained, brittle sandstone		2732	.0073
38	11		Medium-grained sandstone		2743	.0053
39	3	Wabamun	Medium to coarse-grained SANDSTONE		2803	.0071
40	2	Calmar	Shale		2831	.0053
41	3		Dolomitic shale		2857	.0058
42	11	Nisku	Hard quartzitic sandstone		2884	.0086
43	10		Fine-grained sandstone		2896	.0175
44	2		Very coarse sandstone		2915	.0035

The distribution of depths for which samples were obtained is indicative of the principal difficulty in measuring heat flow in Alberta. Core samples from the upper Colorado and from younger formations are relatively rare because the oil industry understandably reserves the expensive coring process for formations not too far removed from the known oil-producing zones.

Birch (1942) gives values for shale conductivity ranging from .0014 to .0066 cal. $sec.^{-1}$ $cm.^{-1}$ $deg.^{-1}$, whereas the conductivities in table 2 for the samples identified as shale range from .0024 to as high as .0153 with an average value of .0063 $\pm$ .0036 (standard deviation). Some of the high conductivities are associated with samples identified as "silty" or "very silty" but others are classified as shales, without qualification. The data quoted by Birch are limited and the more conductive samples in table 2 may represent genuine variations in shale conductivity.

For sandstone samples Birch lists values from .0019 to .0110 which compare somewhat more favoufably with the sandstone conductivities in table 2 which vary from .0031 to .0175 cal. $sec.^{-1}$ $cm.^{-1}$ $deg.^{-1}$.

Again the data quoted by Birch are limited.

6. IMPERIAL LEDUC NO. 503.

6.1 Location and History.

Imperial Leduc No. 503 is located at 53°22' N and 113° 48' W in the valley of the North Saskatchewan River, approximately 20 miles southwest of Edmonton and at an elevation of 2128 feet above sea level (Fig. 5). Drilling of the well began on December 12, 1952 and was completed on February 3, 1953. It was not a producer but has been kept open as a test hole for the use of the oilwell logging companies.

6.2 Temperature Observations, June to September, 1959.

Table 3 lists the temperatures recorded at various depths in Imperial Leduc No. 503, using thermometers A02899, A02906 and AP4028, during the period from June to September, 1959. Also tabulated are the A02899 readings adjusted for the temperature differences observed in the laboratory between the two type AO thermometers, and the type AO results corrected for pressure effects. The A02899 results were adjusted rather than the A02906 because the latter thermometer was used exclusively in Imperial Egremont No. 42. The last column in table 3 gives the average observed differences between the type AP and the type AO readings.

The final temperatures, corrected for pressure, are plotted against depth in figure 6. Also shown is a stratigraphic log derived mainly from the data given in the Schedule of Wells for 1953. The location of the Edmonton-Belly River contact is based on a correlation with the electric log for Imperial Leduc No. 1, located about 9 miles to the southeast.

The best least-squares line through the points was calculated and is shown on the figure. Its slope gives an average geothermal gradient

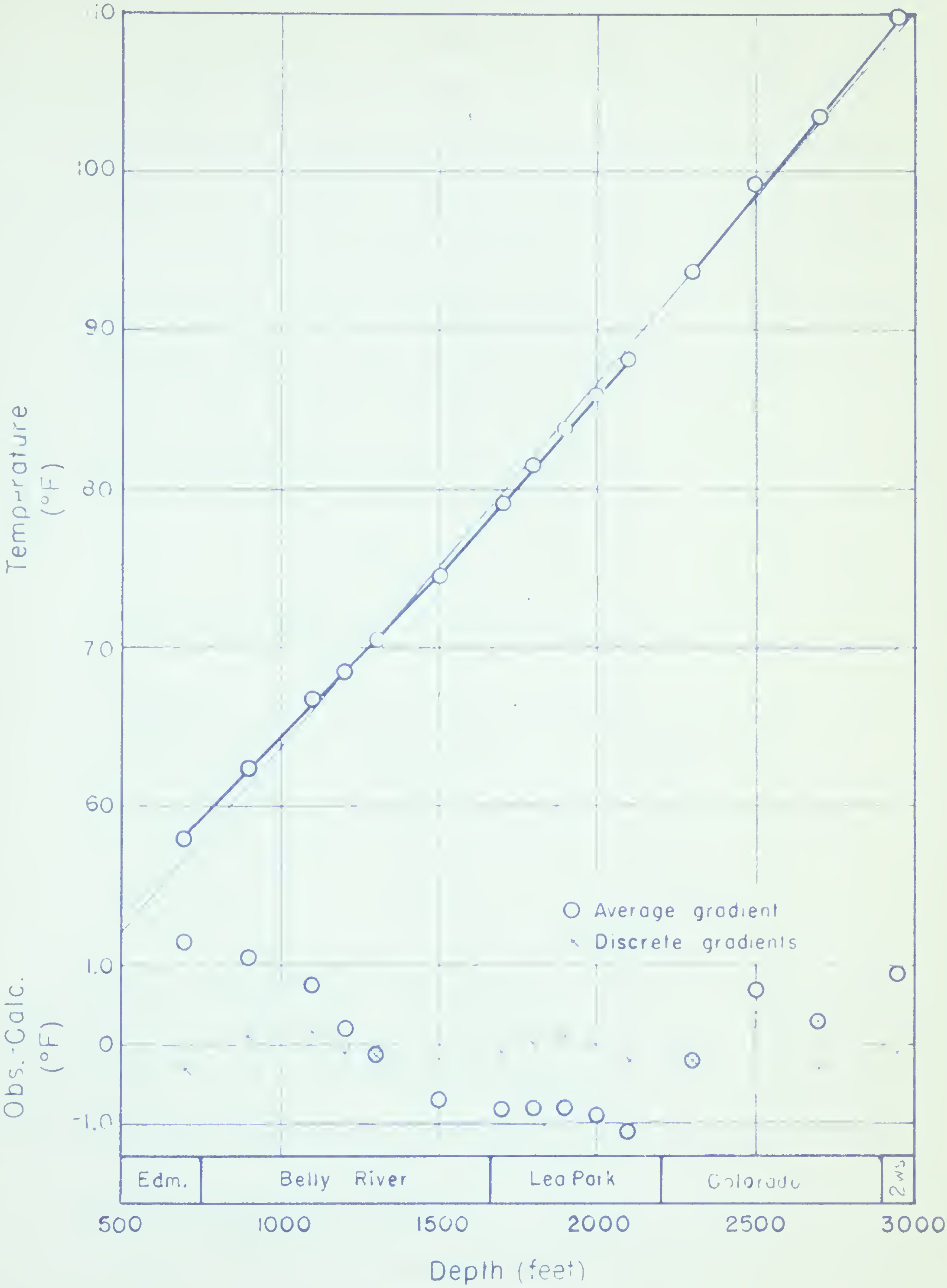
Table 3

Temperatures Recorded in Imperial Leduc No. 503.
June to September, 1959.

Depth (feet)	A02899 (°F)	A02906 (°F) or A02906 Equiv. of A02899 Rdg. (°F)	AP4028 (°F) or Type AO Rdg. Corrected For Pressure	Apparent Temp. Increase (°F), for Type AO, due to Pressure.
500	48.6	48.8	48.8	
700	58.0	58.2	58.0	
900	64.0	64.1	62.5	
1100	69.6	69.7	66.8	
1100	69.6	69.7	66.8	
1100	69.5	69.6	66.7	
1200	72.0	72.1	68.5	
1300	74.7	74.7	70.5	
1300		74.8	70.6	
1500	80.2	80.2	74.6	
1700	86.2	86.1	79.2	
1700	86.0	85.9	79.0	
1800	89.0	88.8	81.5	7.4
1800		89.0	81.5	
1800			81.5	
1900	92.2	92.0	83.8	8.2
2000	94.7	94.5	86.0	8.5
2100	97.8	97.7	88.1	9.6
2300	104.9	104.8	93.7	11.1
2300			93.7	
2500	111.6	111.4	99.2	12.2
2700	117.2	117.2	103.4	13.8
2950	125.8*		109.7	
2950			109.9	
2950			109.8	
2950			109.8	

*Estimate made with aid of rule.

FIGURE 6



TEMPERATURES - IMPERIAL LEDUC NO. 503

for the well of 23.2°F per kilofoot or 42.2°C per kilometer. The differences between observed and calculated temperatures are also plotted. The average deviation of a single temperature reading from the least-squares line is 0.71°F, whereas the average deviation of duplicate readings in table 3 from their means is only 0.04°F. This fact, and the nature of the variation of the observed deviations with depth, suggest that the data can be used to calculate a number of discrete temperature-depth lines, each effective over a limited depth range. These are also plotted on the figure and the corresponding gradients and depth ranges are listed in table 4. The average deviation of a single point is reduced to 0.14°F.

Table 4

Geothermal Gradients in Imperial Leduc No. 503.

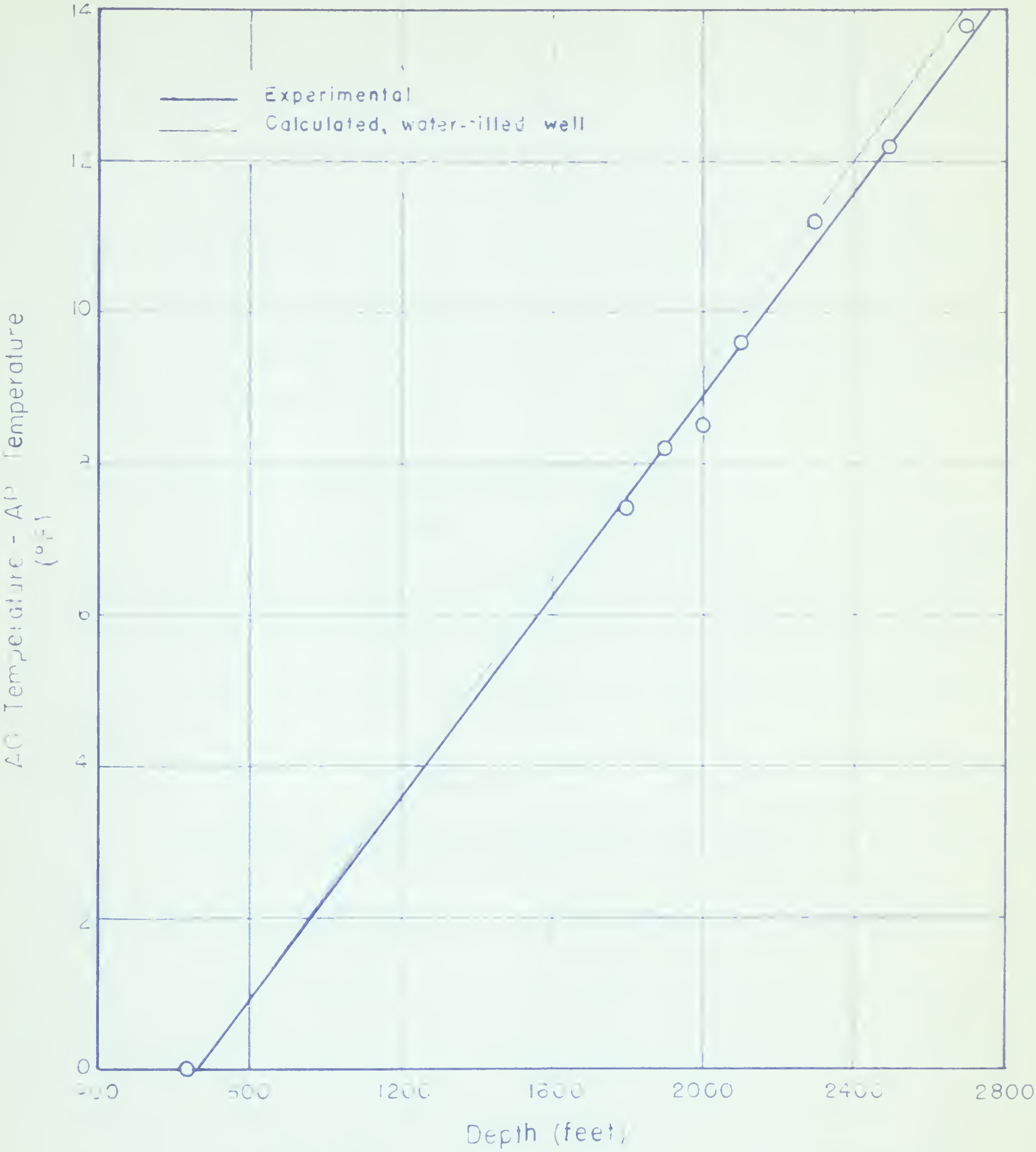
Depth Range (ft.)	Gradient (°F $kft.^{-1}$)	(°C $km.^{-1}$)
700-1500	20.6	37.5
1700-2100	22.6	41.1
2300-2950	24.6	44.8

The three gradients still do not give a complete picture of sub-surface temperature conditions. The 1700-2100 foot and the 2300-2950 foot temperature depth lines do not intersect between 2100 and 2300 feet, indicating that this depth interval may have its own discrete gradient.

6.3 Correction for Pressure Effect.

The data from the last column in table 3 are plotted against depth on figure 7. The fluid level was found to be fairly constant throughout the summer, based on a limited number of observations, and an average

FIGURE 7



PRESSURE EFFECT ON TYPE AO THERMOMETERS
IMPERIAL LEDUC NO. 503

measured value of 636 feet has been adopted. The best straight line through the points is shown, and using the data in Appendix I, the increase in pressure with depth in the well is .413 psi (pounds per square inch) per foot. The corresponding figure for a water-filled well is .434 psi per foot. The actual fluid in the well gave no indication by its appearance that Imperial Leduc No. 503, in this respect at least, was quite similar to a water-filled well. Assumption that the well was water-filled would have increased the corrections for pressure, or decreased the corrected temperatures, by amounts up to 0.3°F.

6.4 Temperature Observations, December 12, 1959.

During the summer months, it was found impossible to obtain readings above certain depths with the maximum thermometers because of the high surface temperatures. An additional trip was made to the well in December so that these readings might be obtained and the results are shown in table 5.

Table 5

Temperatures Recorded in Imperial Leduc No. 503.
December 12, 1959.

Depth (feet)	A02906 (°F)	Depth (feet)	A02906 (°F)
100	33.0	600	46.9
100	36.0	600	47.0
200	37.3	700	48.3
200	37.7	700	46.4
200	37.9	800	60.8
300	40.0	800	60.7
300	40.0	900	63.5
400	41.0	1000	66.1

Table 5

Depth (feet)	A02906 (^{o}F)	Depth (feet)	A02906 (^{o}F)
400	40.2	1100	68.8
400	41.0	2000	94.0
500	42.6		
500	44.5		
500	41.8		

The principal point brought out by the table is the extraordinary variability of the results for depths up to 700 feet. The fluid level had shifted to somewhere between 700 and 800 feet since the series of summer readings and the erratic readings were all taken in the column of air above the fluid. It is obvious that temperatures taken in this column are nearly valueless for heat flow determinations and it is for this reason that the 500-foot result from table 3 has not been plotted on figure 6 or used in the calculations of the gradients.

The effect of the change in fluid level on the apparent temperatures recorded by the type AO thermometers is revealed by a comparison of these temperatures for depths of 900, 1100 and 2000 feet. Those for December 12 average about 0.7^{o}F less than the summer readings, indicating, according to figure 7, a drop of slightly more than 100 feet in the fluid level. The observed temperature differences cannot be attributed to the different thermometers used because they gave almost the same readings during the summer. Furthermore, although no attempt was made to definitely locate the fluid level in December, there was no doubt that it lay between 700 and 800 feet because the thermometer came out of the well dry until readings were made at 800 feet.

7. IMPERIAL EGREMONT NO. 42

7.1 Location and History.

Imperial Egremont No. 42 is located in a creek valley at 54°00' and 113°07' W, about 3 miles north of the town of Redwater and 35 miles north-east of Edmonton (Fig. 5). The elevation above sea level is 2032 feet. Drilling of the well began on March 22, 1950 and was completed on March 30. The hole was lined with 7" casing (O.D.), fitted with 2½" central tubing, and was placed on production on April 5. Production was from the Leduc formation, the top of which is 3112 feet below the surface, and, for production purposes, the hole was left uncased from 3117 to 3158 feet.

During the production period, 73,125 barrels of oil were pumped from the well. The average specific gravity of the oil produced was 0.85 and for each barrel of oil, 184 cubic feet of gas were evolved. In October, 1952, the production period ended and the well was shut down for conversion into a pressure observation well. Since that time, pressures at various depths in the well have been measured at three-month intervals by the Oil and Gas Conservation Board in Redwater. Wellhead pressures vary up to a maximum of about 40-50 psi.

7.2 Temperature Observations.

Table 6 lists the temperatures recorded in Imperial Egremont No. 42, using thermometers A02906 and AP4028, during November and December, 1959. Also tabulated are the type AO readings corrected for pressure, and the observed average difference between the type AP and type AO results. In making these observations it was generally necessary to use a device known as a lubricator (Fig. 8) on the wellhead to reduce to a minimum the release of gas from the well, the lubricator being essentially a pressure-tight cylinder large enough to contain the probe. The wellhead valve was only

Table 6

Temperatures Recorded in Imperial Egremont No. 42.
November and December, 1959.

Depth (feet)	A02906 (°F)	AP4028 (°F) or Type AO Rdg. Corrected For Pressure	Apparent Temp. Increase (°F), for type AO, due to Pressure
888	54.4	54.4	
890.5	54.4	54.4	
900	54.3	54.2	
1000	57.0	56.3	
1000	57.4	56.7	
1000	57.5	56.8	
1000	57.4	56.7	
1200	63.0	61.1	
1200	63.3	61.4	
1205	62.9	60.9	
1205	63.0	61.0	
1400	67.2	64.1	
1505	69.9	66.2	
1605	72.8	68.5	
1800	78.2	72.8	
2005	84.2	77.7	
2005	84.2	77.7	
2105	86.9	79.8*	7.1
2205	89.3	81.8	7.5
2305	91.9	83.8	8.1
2500	95.3	87.5	8.1
2500	95.6		
2505		87.3	
2705	99.8	89.7	10.1
2955	103.3	92.1	11.3
2955	103.2	91.8	

* Estimate.

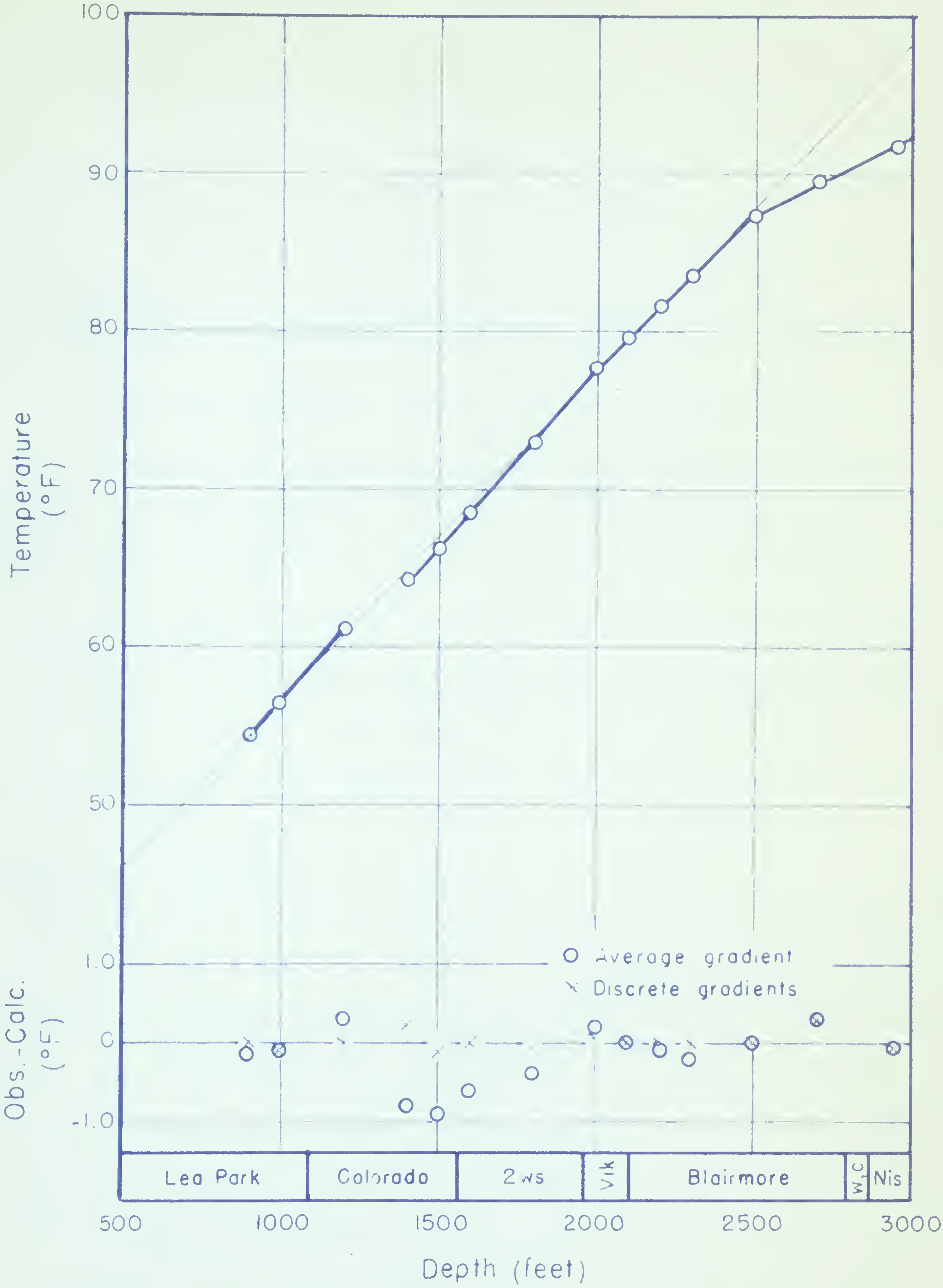
FIGURE 8

LUBRICATOR

opened after the probe had been sealed within the lubricator and was shut off before the probe was removed. On occasion, however, the gas within the well was inadvertently brought to atmospheric pressure and, when this occurred, the lubricator was inconvenient to use and served no useful purpose in any case. The depths in table 6 which are five feet beyond the even hundred result from correcting the depths without lubricator to the datum effective for measurements with the lubricator.

The final corrected temperatures are plotted against depth on figure 9, together with a stratigraphic log taken from the Schedule of Wells for 1950. A striking difference between the results for this well and those for the Leduc well, is the existence of an abrupt change of geothermal gradient at a depth of about 2500 feet. Best least-squares lines are shown for the depth ranges above and below this discontinuity. In the upper part of the well the average gradient is 21.0°F per kilofoot (38.2°C per km.) and in the lower 10.0°F per kilofoot (18.2°C per km.). Deviations of the individual temperatures from the two lines are also plotted against depth on the figure and again the existence of a number of minor changes in geothermal gradient may be deduced. The corresponding temperature-depth lines are plotted on the figure, and the gradients and depth ranges given in table 7. With this set of minor changes the average deviation of a single point is reduced to 0.12 from 0.26°F. The average deviation of duplicate readings in table 6 from their means is 0.07°F.

FIGURE 9



TEMPERATURES - IMPERIAL EGREMONT NO. 42

Table 7

Geothermal Gradients in Imperial Egremont No. 42

Depth Range (ft.)	Gradient ($^{o}F\ kft.^{-1}$)	($^{o}C\ km.^{-1}$)
888-1205	22.6	41.1
1400-2005	22.8	41.5
2005-2470	20.0	36.4
2470-2955	10.0	18.2

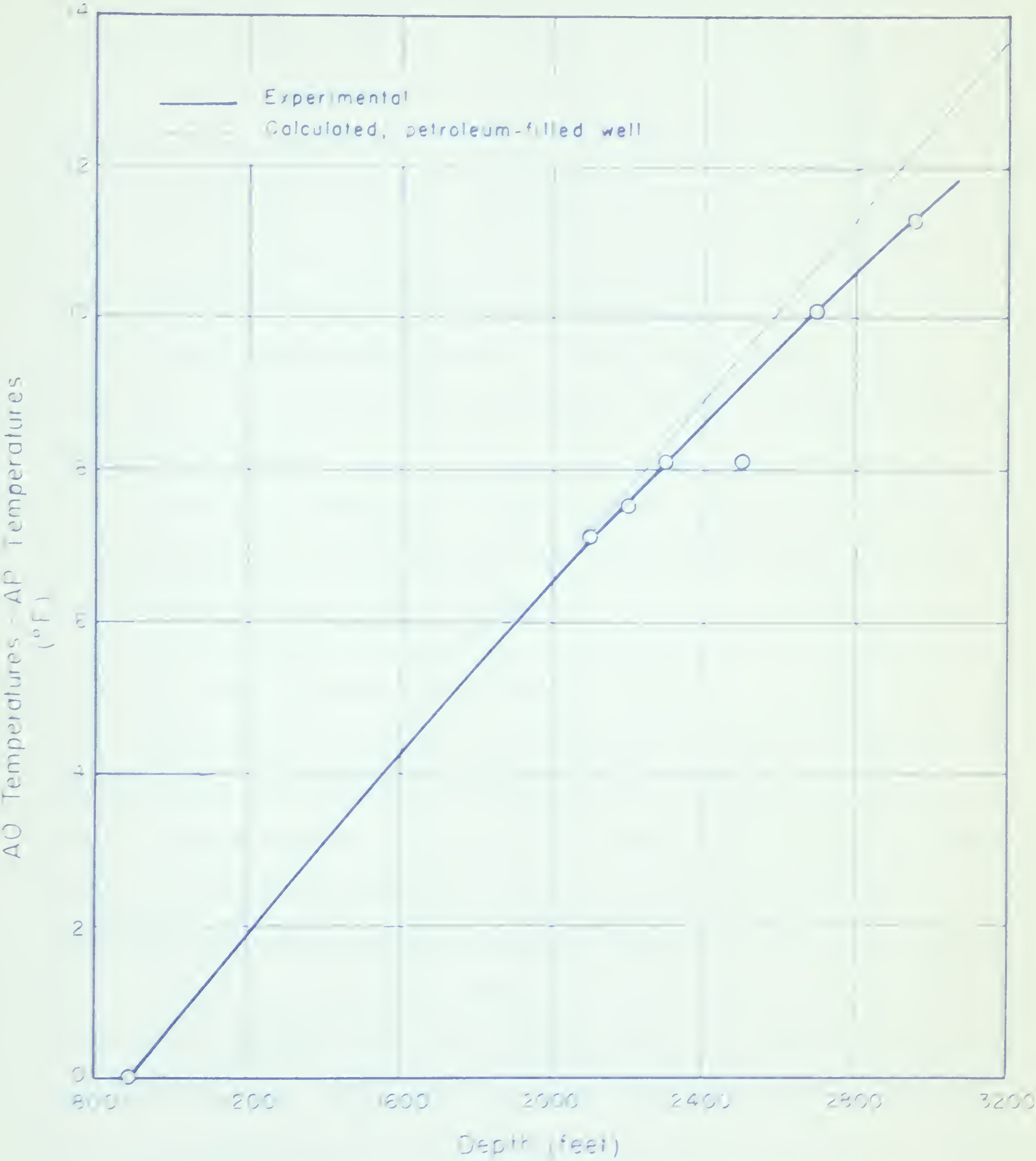
As in the case of Imperial Leduc No. 503, this picture of the subsurface temperature conditions is incomplete. The 888-1205 foot and the 1400-2005 foot temperature-depth lines do not intersect between 1205 feet and 1400 feet, so that another discrete gradient must exist within this range.

In addition to the temperatures in table 6, a number of others were obtained in the gas column. Results were similar to those for the Leduc well given in table 5.

7.3 Correction for Pressure Effect.

The data from the last column in table 6 are plotted against depth on figure 10. The fluid level again appeared to be reasonably static during the period over which temperatures were measured, and an accurate determination placed it at 887 feet. The points, except for the anomalous one at 2500 feet, fit a parabolic relation much better than they do a linear one. Above 2000 feet the parabola closely coincides with the calculated straight line for a well containing fluid with a density of 0.85. On this basis there is some reason to suspect a pressure disturbance at depth within the well. Temperature readings at 2500 feet were repeated using both types of thermometers, and there seems to be no doubt that this point lies off the parabola. If there is fluid flow in the well, the point of inflow may be close to 2500 feet,

FIGURE 10



PRESSURE EFFECT ON TYPE AO THERMOMETERS
IMPERIAL EGREMONT NO. 42

causing an increased pressure drop there. The calculated pressure drop at 2500 feet is 62 psi. The abrupt gradient change at 2470 feet may also be a symptom of fluid flow and all the evidence presented would suggest flow into the well at about 2500 feet and from thence down toward the producing formation. It is only fair to point out that Imperial Oil officials at Redwater do not consider this to be likely; that the pressure gradient as measured by the Oil and Gas Conservation Board appears to be completely normal and that for such an effect to be observable both the outer casing and the central tubing must be leaking. Nevertheless, there is some justification for treating the temperatures for depths over 2500 feet with some degree of caution.

An empirical correction was again used to adjust the type AO results for pressure effects, the correction being based on the parabolic relation between apparent temperature and depth.

7.4 Temperature Disturbance due to Well Production.

Use of the data in section 7.1 concerning the well's production history enables the temperature disturbances due to fluid flow and gas expansion during production to be calculated from equations 7 and 16 respectively. The residual temperature disturbance at the time temperature observations were made can then be computed from equation 19.

For the case of gaseous expansion it is assumed that all the expansion to atmospheric pressure takes place adiabatically at the point where the gas flows into the well. The average mass rate of gas flow into the well over the period of production, assuming the gas to be methane only, was 3.48 grams per second. When this mass of gas expands on entering the well, it cools, and heat is removed from the neighbouring formations to warm it. The heat

that must be supplied in this way to balance the cooling of the gas is 414 calories per second, or 1.66 calories per second per square centimeter over the cross-sectional area of the well. This is the quantity q in equation 16.

Other constants required for the calculations of the temperature disturbances are the following:

$k = .007$ cal. sec.$^{-1}$ cm.$^{-1}$ deg.$^{-1}$

$\rho = 2.50$ gm. cm.$^{-3}$

$c = 0.18$ cal. gm.$^{-1}$ deg.$^{-1}$

$\kappa = .0156$ cm.^{2} sec.$^{-1}$

$a = 8.9$ cm.

$t = 7.86 \times 10^{7}$ sec.

$V = 111$ cm.^{3} sec.$^{-1}$

$\rho_f = 0.85$ gm. cm.$^{-3}$

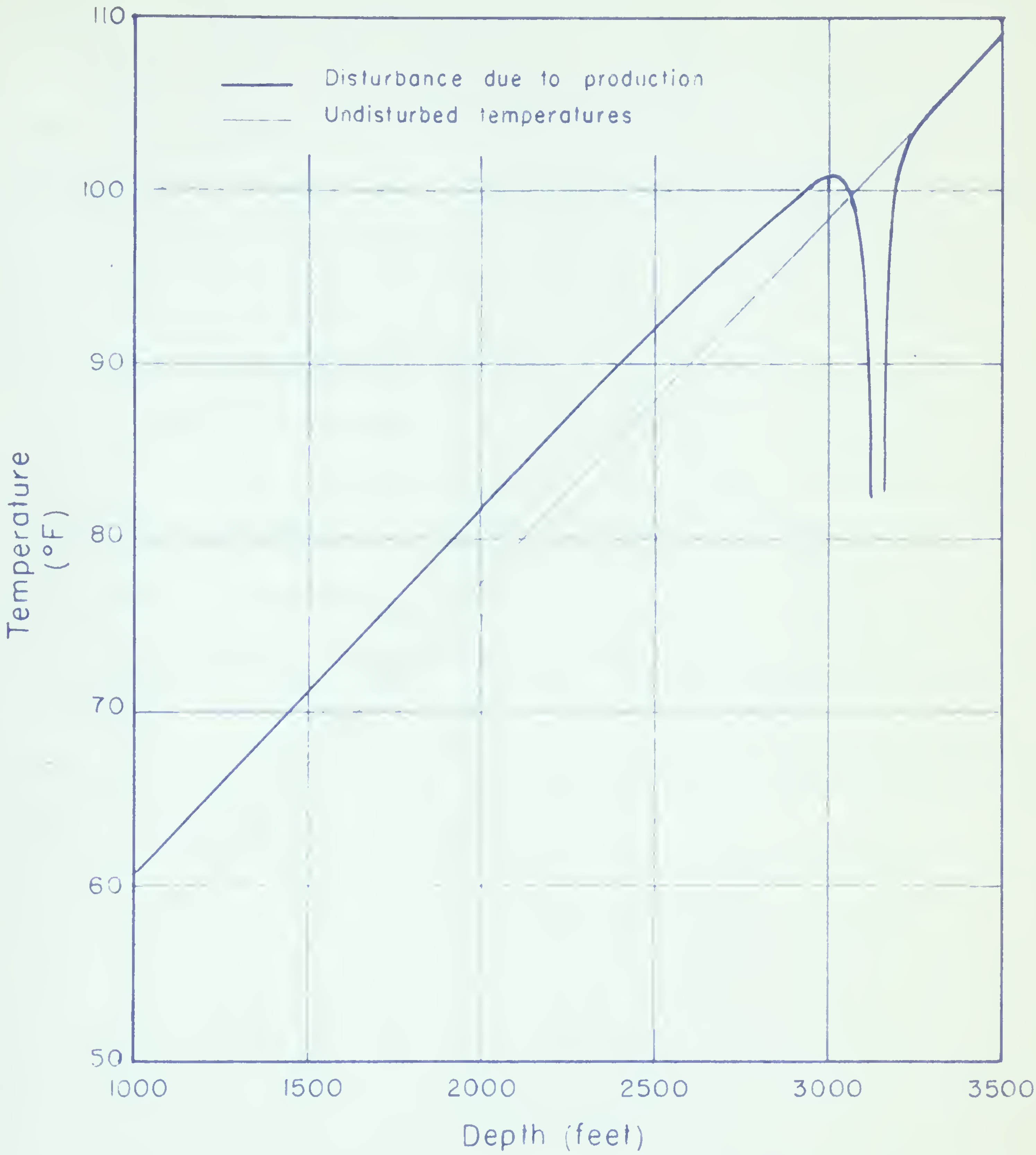
$c_f = .511$ cal. gm.$^{-1}$ deg.$^{-1}$

$g = 3.82 \times 10^{-4}$ $^{\circ}$C cm.$^{-1}$

The values for k and g are experimental; ρ, c and c_f are values taken from tables, the first two being representative of a number of earth materials; and the other quantities are derived from the data in section 7. 1.

The calculated combined temperature disturbance for fluid flow in the well, and for gaseous expansion, as they were at the time production stopped are shown on figure 11. That due to fluid flow increases fairly rapidly to a maximum of 4.0°F greater than the undisturbed temperature. The disturbance due to gaseous expansion is very large in the immediate vicinity of the inflow point, but becomes negligible within 150 feet. The actual effect was in all probability much smaller since all the expansion did not take place at the point of inflow, but only that due to the change

FIGURE 11

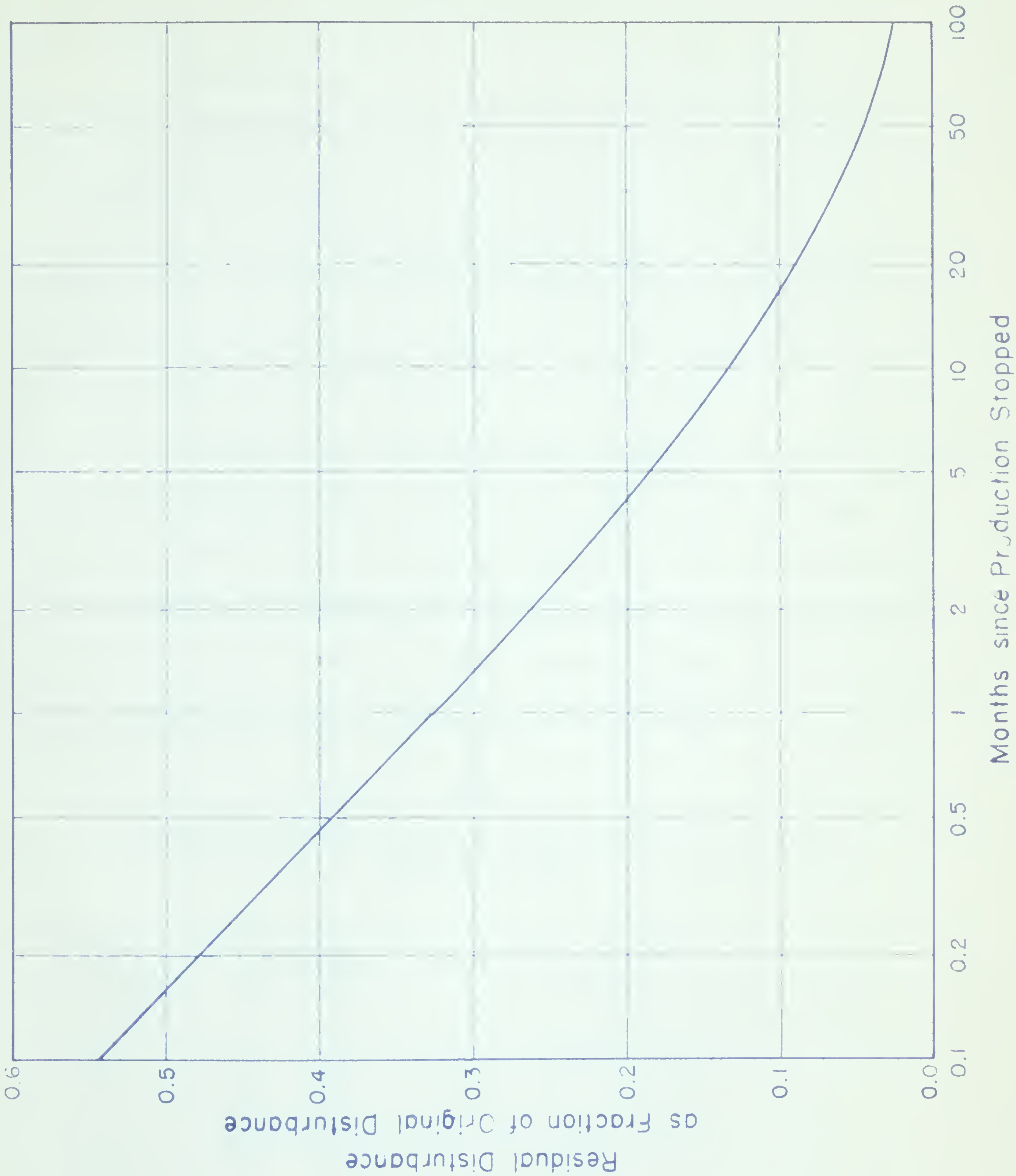


TEMPERATURE DISTURBANCE AT END OF PRODUCTION PERIOD
IMPERIAL EGREMONT NO. 42

from formation pressure to hydrostatic pressure. In the Redwater field the difference between these two pressures is small.

Considering the other extreme of the gas expanding gradually as it rises to the surface, 414 calories per second must be supplied from the surroundings by the rising fluid. From this point of view, the calculated maximum temperature disturbance due to fluid flow is decreased by about 24 percent, to 3.1 F. However the overall effect is still a localized heating in the neighbourhood of the well.

The residual disturbance at the time temperatures were measured in the well can now be computed from equation 19. Figure 12 shows a graph of the residual disturbance as a fraction of the original disturbance, plotted against the time since the well was shut down. For 86 months, which is the appropriate interval for the measurements described here, the disturbance has been reduced to 0.029 of its original value. The maximum disturbance is now only 0.1^{o}F and the formations surrounding the well have virtually regained thermal equilibrium.



RETURN TO THERMAL EQUILIBRIUM
IMPERIAL EGREMONT NO. 42

CALCULATED HEAT FLOW

8.1 Imperial Egremont No. 42.

For the calculation of heat flow in Imperial Egremont No. 42, the data on thermal conductivity and equivalent depth in table 2 were combined to calculate $\sum t_i/k_i$.

It was assumed that bed thicknesses were defined by depths midway between those quoted in the table. For instance, sample no. 20, with a conductivity of .0044 cal. sec.$^{-1}$ cm.$^{-1}$ deg.$^{-1}$, was considdred to represent a bed extending over a depth range from 2309 to 2398 feet. Because there is only one conductivity above 1835 feet, and this is considerably removed from the others, the origin for $\sum t_i/k_i$ was placed at 1800 feet. The table developed from these calculations gave $\sum t_i/k_i$ for each depth at which it had been assumed a conductivity change occurred, and from this table the values for depths at which temperatures had been measured were also computed. The final results are shown in table 8.

Table 8

Temperatures and Calculated Values of $\sum t_i/k_i$ in Imperial Egremont No. 42.

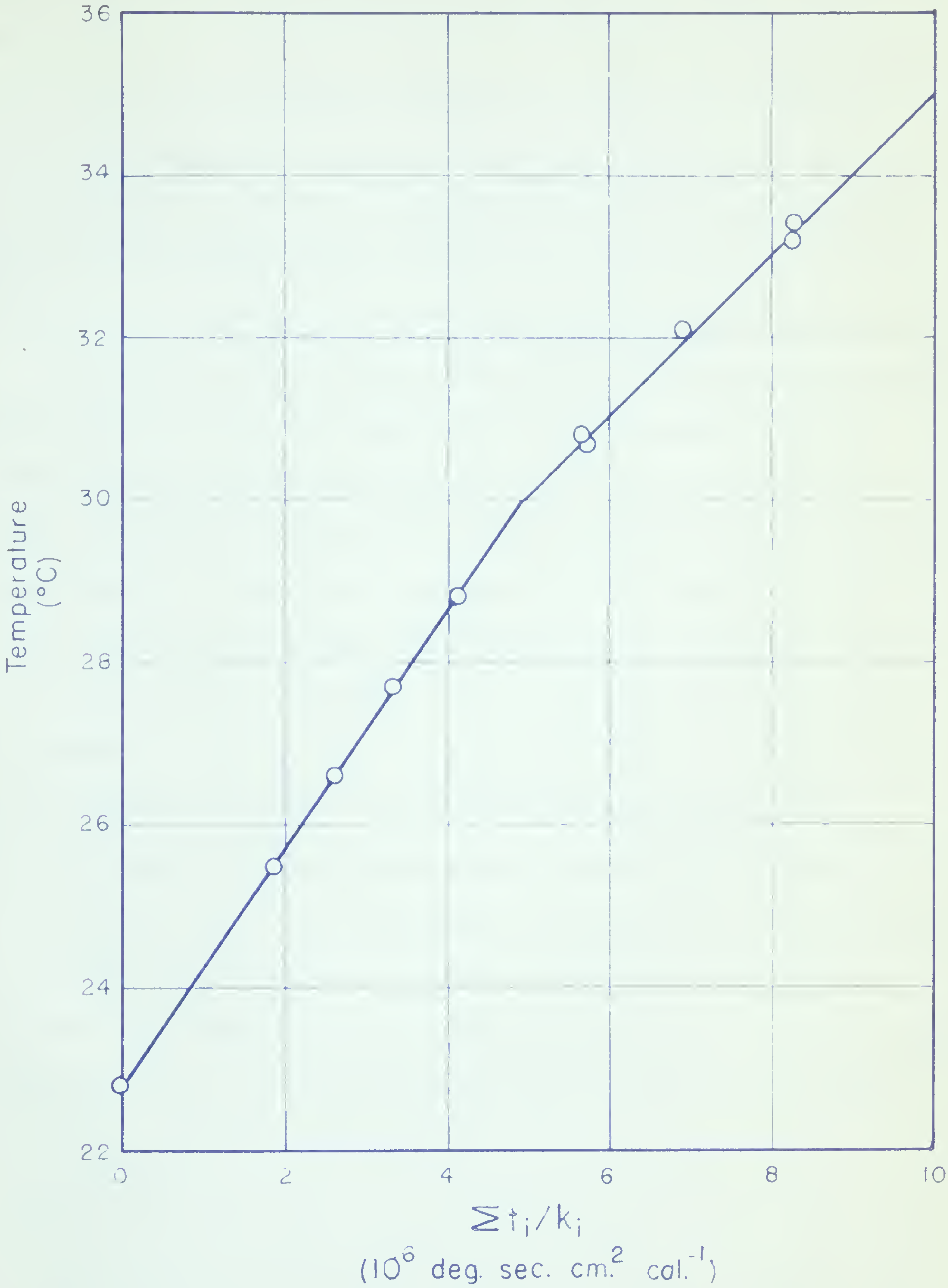
Depth (feet)	Temperature (^{o}F)	(^{o}C)	t_i/k_i Cal. -1 cm.^{2} sec. deg.
1800	73.0	22.8	0.00×10^6
2005	77.9	25.5	1.88
2005	77.9	25.5	1.88
2105	79.8	26.6	2.61
2205	81.8	27.7	3.33
2305	83.8	28.8	4.14
2500	87.5	30.8	5.68
2505	87.3	30.7	5.73
2705	89.7	32.1	6.94
2955	92.1	33.4	8.28
2955	91.8	33.2	8.28

The graph of temperature against $\sum t_i/k_i$ is shown on figure 13. The result is not the single straight line predicted by equation 5 for steady-state flow in a medium with horizontal, homogeneous and isotropic layering, but instead appears to be best approximated to by a pair of straight lines, as shown in the figure. Their slopes are 1.46 ± 0.004 and $0.99 \pm .05 \times 10^{-6}$ cal. cm.$^{-2}$ sec.$^{-1}$. The two lines intersect at $\sum t_i/k_i = 1.6 \times 10^6$, equivalent to a depth of about 2410 feet. This compares with the calculated depth of 2470 feet for the change in geothermal gradient from 36.4 to 18.2°C km.$^{-1}$.

It is evident that some of the gradient change must be attributed to an increase in the average formation conductivity. The change in gradient at 2470 feet is a decrease of 50 percent, whereas the comparable decrease in slope at 2410 feet is only 32 percent. Determination of the average conductivity above and below 2470 feet gives $.0049 \pm .0026$ for the upper part of the well and $.0080 \pm .0040$ for the lower part. If Student's t-test (Cameron, 1955) is used to test the significance of the difference between these two averages, t is found to be 2.94 for 41 degrees of freedom. This value of t is significant at the one percent level; that is it would be duplicated or exceeded only one percent of the time in random drawings of similarly-sized samples from the same population. There is, therefore, a 99 percent chance that the two are drawn from different populations, or that the difference between the two average conductivities is significant. Thus the conductivity change is certainly a contributory factor to the change in gradient, but is not, in itself, large enough to completely explain it.

The separation of samples into groups according to depth in the well is much more significant statistically than the previous separation that was made on the basis of lithology (Section 5.4). Nevertheless, the higher

FIGURE 13



HEAT FLOW DETERMINATION
IMPERIAL EGREMONT NO. 42

average conductivity from 2470 feet to 2955 feet may possibly reflect lithologic changes. The data in table 1 indicate that once the glauconitic-ostracod series in the Blairmore formation is reached, the formations contain greater proportions of sand and other more highly-conducting materials, such as limestone and dolomite. According to data on the Egremont well supplied by Imperial Oil Limited, the top of the glauconitic sand is at 2551 feet.

If a single straight line had been obtained for the relation between temperature and $\sum t_i/k_i$, its slope, according to equation 5, would unequivocally determine the geothermal heat flow in the neighbourhood of this well. The fact that there are actually two straight lines introduces some uncertainty, but the doubts already cast on the validity of the temperature measurements at depths of 2500 feet and more suggests that the second line be rejected. Taking the relative standard deviation in the brass conductivity into account, $1.46 \pm 0.08 \times 10^{-6}$ cal. cm.$^{-2}$ sec.$^{-1}$ is therefore accepted as the true value of the heat flow for this well. Whether this value can be permanently established depends on future work in the Redwater field. If the explanation advanced for the anomalous depth range is correct, it is unlikely that such a region will be met with in other wells, especially in the same stratigraphic position. If such a region is encountered again, the explanation for it and the heat flow estimate above would both have to be reconsidered.

8.2 Imperial Leduc No. 503.

Determination of terrestrial heat flow in Imperial Leduc No. 503 is semi-qualitative at best because there is only one measured conductivity which can be used in this well. The calculated depth at which it applies is 2264 feet, and, unfortunately, this is in the depth interval in which

the gradient is not accurately determined. Limits can be set, however, between which the gradient probably lies. These are 44.8 and 54.2°C km.$^{-1}$, which are, respectively, the slope of the 2300- to 2950- foot temperature-depth line and the slope of a line drawn through the points from 2100 to 2500 feet. Using these gradients and .0037 cal. sec.$^{-1}$ deg.$^{-1}$ for the conductivity, the heat flow is calculated to be in the range 1.7 to 2.0 $\times 10^{-6}$ cal. cm.$^{-2}$ sec.$^{-1}$. The conductivity used is low compared to the average of .0049 for somewhat deeper formations, and if the latter should turn out to be more representative of the conductivity of the Colorado group, the heat flow estimate would have to be increased.

An alternative method of obtaining an approximate heat flow value is to compare the gradients in similar formations for the Leduc and Redwater wells. If the conductivity of all formations remains unchanged between the two wells, the gradients in one will be proportional to those in the other, the ratio of proportionality being the same as that for the two heat flows. If proportionality is not observed, a number of heat flow estimates may be made by considering each formation in turn to have its conductivity unchanged. A list of conductivities for similar formations is given in table 9.

Table 9

Comparable Geothermal Gradients in the Leduc and Redwater Wells.

Imperial Leduc No. 503 (°C km.$^{-1}$)	Imperial Egremont No. 42 (°C km.$^{-1}$)
41.1	41.1
44.8-54.2 (?)	28.9 (?)
44.8	41.5

It is apparent from the table that at least two of the formations suffer changes in average conductivity between Leduc and Redwater. From the ration of the gradients in each formation, and from the calculated heat flow for Imperial Egremont No. 42, the heat flow for the Leduc well can be estimated to be in the range 1.5 to 2.7 x 10^{-6} cal. cm.$^{-2}$ sec.$^{-1}$. If the rather questionable gradients for the intermediate formation are neglected, the range is reduced to 1.5 to 1.6 x 10^{-6}.

Although the determination of an accurate heat flow for Imperial Leduc No. 503 cannot be made until considerably more conductivity data becomes available, it seems fairly certain that the flow in this well is as high as, or higher than, that in the Redwater well. If this is indeed the case, the explanation may lie in fundamental differences in the radioactive content of the basement rocks underlying the two regions. Garland and Burwash (1959), on the basis of evidence from gravity measurements and petrological analysis of cores from wells penetrating the basement, have prepared a map indicating the probable basement lithology for the area of central Alberta. The three lithologic types mapped are gneissic, granitic and basic. According to the map the Redwater well overlies basic rock, in the neighbourhood of a basic rock to gneiss contact. The Leduc well is over gneiss, somewhat further removed from a granitic rock to gneiss contact. The differences in radioactive heat production, according to Bullard (1954), between sialic and simatic rocks have already been mentioned. These two terms are synonymous with granitic and basic, so that local variations in heat flow could easily arise because of the local variations in basement lithology.

8.3 Discussion.

The major cause of imprecision in the final result apparently originated in the calibration of the brass bars. The calculated standard deviation for the slope of the line in figure 13 which defines the heat flow for the Egremont well was negligible. This result may have been purely fortuitous and it does not seem likely that it will be duplicated. Nevertheless, it may indicate that the individual errors in conductivity tend to cancel out when the results are combined in the manner suggested by equation 5. It has already been demonstrated that there is a high degree of reproducibility for the maximum thermometer readings. The standard deviation for the brass conductivity on the other hand is about 5 percent. If future results indicate that the precision of the determination of the slope from which the heat flow is deduced remains high, it would probably be worth while to investigate ways of improving the precision of the conductivity measurement, at least in the case of the brass bars.

If the scope of heat flow determinations in Alberta is to be increased, and if useful correlations are to be made between different areas, emphasis should be placed on a systematic search for core samples from such formations as the Colorado group, and others higher in the stratigraphic column. Though such samples may be rare, five to ten evenly spaced over a thousand feet could be sufficient to provide a good heat flow estimate in wells such as Imperial Leduc No. 503.

Until such samples are available, future temperature measurements should be confined to depth ranges and wells for which the existing conductivity data is applicable, at least until the faster logging method has been rendered reliable. In this way a maximum of information can be obtained

from a minimum of temperature data. A great percentage of the temperatures reported in this investigation were of no immediate value because of the lack of corresponding conductivities.

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APPENDIX I

Laboratory Calibration of Thermometer A02906 for Pressure.

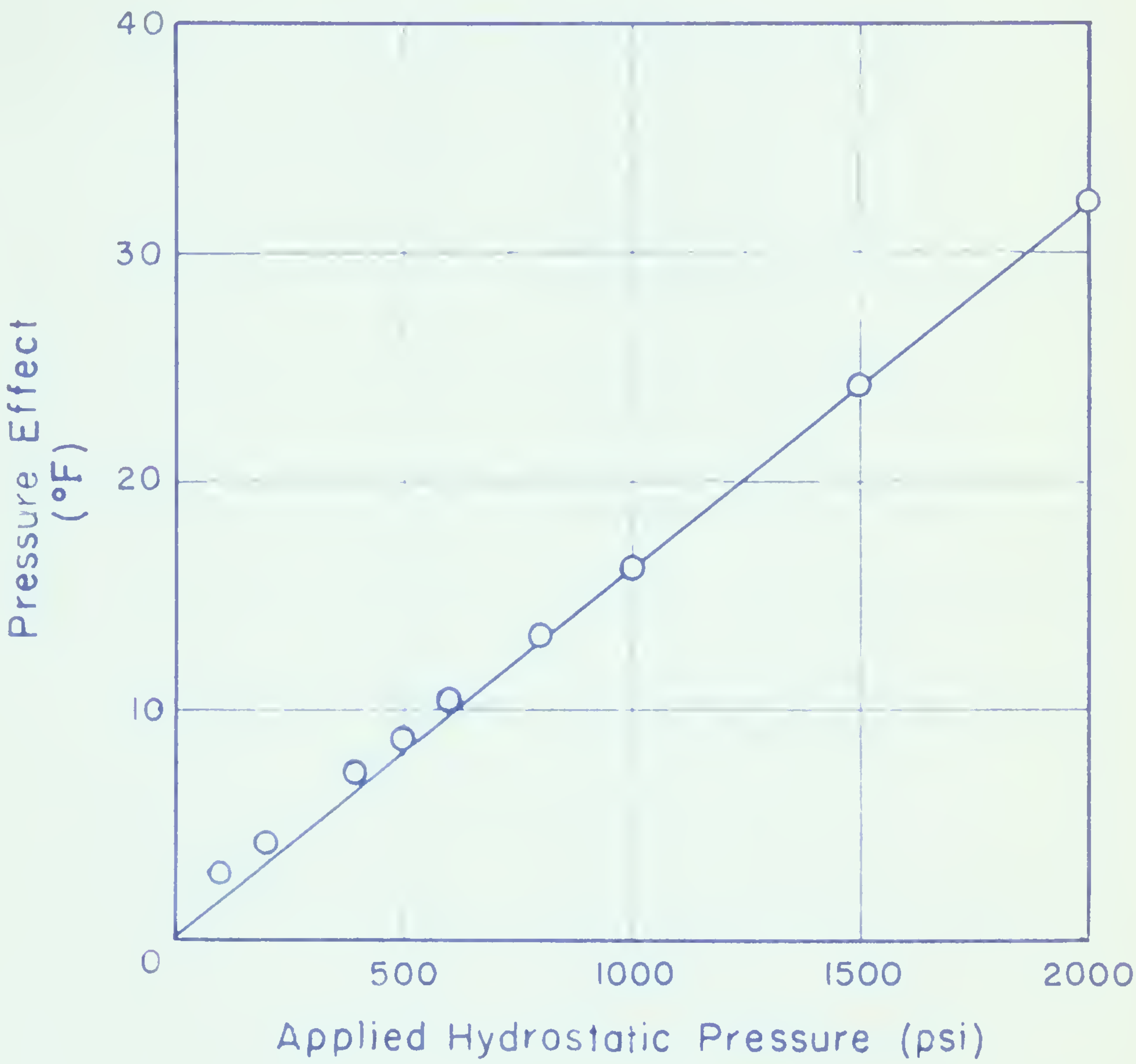
Applied Pressure (psi)	Pressure Effect (°F)	Applied Pressure (psi)	Pressure Effect (°F)
100	2.9	1500	24.0
200	4.1	1500	24.1
400	7.2	2000	32.4
500	8.8	2000	32.1
500	8.9	2000	32.2
500	8.8	2000	32.2
500	8.8	2750	43.0
600	10.3	2750	43.1
800	13.2	2750*	43.3
1000	16.2	2750*	42.7
1000	16.1		
1000	16.1		
1000	16.3		

* For these two results, the apparatus was immersed in an ice bath. For all the others the apparatus was at room temperature.

The results of the calibration are plotted in figure 14. The results for 2750 psi are not included because they are based on readings with a different pressure gauge.

The reason for the failure of the experimental points at low pressures to fall on the straight line through the origin was not investigated. In any calculations based on this data it has been assumed that the straight line through the origin defined by the points for 1000, 1500 and 2000 psi determined the pressure effect.

FIGURE 14



CALIBRATION OF A02906 FOR PRESSURE

APPENDIX II

List of Cored Wells

Well Number	Well Name
1	Imperial Egremont 20
2	Imperial Redwater 21
3	Imperial Redwater 1
4	Imperial Redwater 11
5	Imperial Redwater 50
6	Imperial Amelia 24
7	Imperial Amelia 57
8	Northern L.P. Mann No. 1-23
9	Imperial Simmons 103
10	Redwater Petroleums 1
11	Kozak 16 - 3
12	Sun et al Calling Lake Province 1
13	Anglo-Home- C & E Elk Point 3
14	Anglo-Home- C & E Elk Point 1A